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REFERENCE

Lane, B. H., 1935. Suggestions to authors of papers submitted for publication by the United States Geological Survey: U. S. Geol. Survey, Washington, D. C. (For sale by Superintendent of Documents, Washington, D. C., 25c.)

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JANUARY 1952

SEDIMENTATION AND STRATIGRAPHY OF THE HURONIAN OF UPPER MICHIGAN

S. A. TYLER and W. H. TWENHOFEL

PART 1

ABSTRACT. The Huronian section of the Upper Peninsula of Michigan, particularly the section in the Marquette district, is described in detail and the writers' views relating to the conditions and environments of deposition are presented. The conclusion is reached that the several lithic units of each major division are to some degree lateral equivalents and are not completely sequential as seems to have been generally assumed.

The unconformity postulated in the Marquette district between the assumed Lower and Middle Huronian is not recognized and is believed not to exist. The features cited by previous geologists as evidences of an unconformity are believed to be the result of deformation.

The different lithic units have had different environments of deposition and only one lithic assemblage—the Kona-Randville-Bad River dolomitic limestones—is considered to have been deposited under marine conditions. The other lithic units, and this includes the iron formations, are believed to have been deposited under deltaic or some related environments. No lithic unit is a time unit.

INTRODUCTION

ALTHOUGH the Huronian formations of Upper Michigan have been studied for the past 100 years, many problems relating to the origin and the conditions existing at the times of deposition remain unsolved. The association of chert and iron-bearing minerals, unique to the Precambrian, has not been adequately explained and the development of animal and plant life during the Huronian is essentially unknown.

Whenever and wherever sediments are deposited, gradations of gravels into sands, sands into muds, and muds into lime sediments may be observed, yet the literature of the Precambrian shows that this elementary basic fact of sedimentation has not been applied. The long-discredited Wernerian doctrine of sedimentation which assumed uniform marine

conditions of world-wide extent still seems to influence the thinking of many students of the Precambrian. This is exemplified by the dogged insistence that similar lithic units exposed at widely separated localities represent the same span of geologic time, or that similarity of lithic sequence automatically means equivalence in geologic time.

This paper is an attempt to unravel the depositional history of the Huronian rocks of Upper Michigan in the light of modern concepts of sedimentation with the hope that this form of approach may aid in solutions of some of the problems. The problem is difficult in view of the fact that over large areas the rocks are highly deformed and metamorphosed, outcrops are relatively rare in most districts because of thick glacial drift, and lateral gradations from one lithology to another are rarely or poorly exposed.

The Marquette district was selected for careful study because the deformation and metamorphism are rather mild in the eastern part of that area and rock exposures are relatively numerous, but Huronian sections were also examined in the Menominee, Crystal Falls, Penokee-Gogebic and Iron River districts in order to ascertain the regional extent of lithic horizons and the degree and character of facies changes.

This study was financially supported by a grant from the Jones & Laughlin Steel Corporation. The authors are especially indebted to Mr. L. P. Barrett, chief geologist of the above company, whose great interest in the Precambrian made the study possible. Thanks are also due to Mr. Emil Kronquist and Mr. Edward Buetner for information and to Mr. Philip Oetking who assisted in the field.

GENERAL PRINCIPLES OF SEDIMENTATION

Any study of the geologic history of the Precambrian rocks must accept the concept that basic principles of sedimentation provide the framework for and the key to the interpretation of their depositional history. Since the general character of the Huronian sediments indicates deposition in, or in some cases perhaps adjacent to standing bodies of water, the following principles of sedimentation must be given consideration: (1) physical conditions controlling deposition of sediments, (2) relationship of the size, shape, and depth of the depositional basin to the type of sediment and thickness of each

lithic unit, (3) relations of lithic units to time units, (4) influence of relief and vegetation in source areas and of distance of transportation upon the character of sediment deposited, (5) relation between atmospheric composition and manner of transportation and deposition of iron compounds, (6) formation and preservation of excellent lamination.

Physical conditions controlling deposition of Huronian sediments.—The deposits of the Huronian accumulated over areas undergoing submergence. It is probable that the areas of deposition were isolated at first and that as submergence progressed the isolated areas gradually merged to form extensive bodies of water. The basal sediments spread laterally and at the same time became later in time. Uplift or filling of the submerged area terminated a cycle.

If Walcott's (1910, pp. 1-16) view that the large land areas of Precambrian time had much the same outline as at present be accepted, northern Michigan was far in the interior of the North American continent and was far from the sea. The Huronian areas of deposition may at times have been connected during each cycle of their existence with the open

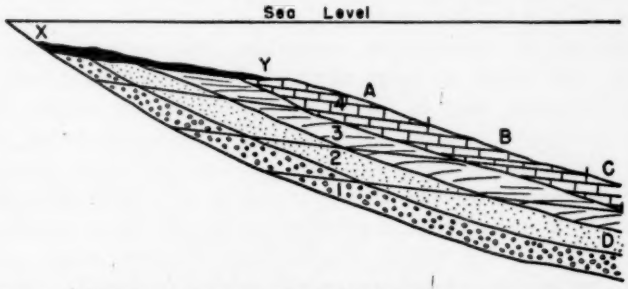

Fig. 1. Diagram showing theoretical distribution of sediments made by advancing waters. 1, 2, 3, and 4 represent lithic units; A, B, C, and D time units. The lithic units are more steeply inclined than the time units. Lithic units 1 and 2 may be considered to represent the Mesnard quartzite. It should not be assumed that the lithic units are as sharply defined as the diagram indicates. In each standstill position of water level, sediments like those of 1 and 2 would be extended basinward and to a greater or less degree would overlie sediments 3 and 4 as indicated by the heavy line which indicates base level of deposition. The diagram shows the lithic units have somewhat uniform thickness. There would probably be considerable variations in thickness. Extensions of units 1 and 2 at standstill position might represent the Ajibic quartzite which on advance of the waters would be extended outward from the basin.

sea, but most of the time probably had no connection therewith. Quirke and Collins (1930, p. 343) express the view that the Grenville limestones of eastern Ontario, Quebec, and the Adirondack region may be the metamorphosed marine equivalent of the Huronian. Leith (1934, p. 178) shares the same view, for he states, "I find it more plausible to assume that the Grenville represents the marine, or at least water-deposited, equivalents of the partly continental and partly marine sediments of the Huronian area."

The extent of connection of the Huronian area of Upper Michigan with the open sea must have exerted profound influence upon sedimentation within the basin or basins. If connection was relatively wide, sediments would accumulate within the basin or basins until the entire bottom had been built to the base level of deposition after which sediments would tend to be transported to the open sea. The situation would resemble that which exists today in Hudson Bay. If there were very restricted connections with the open sea, all sediments would be likely to remain in the area and the basin or basins could become filled.

The salinity of the waters over Upper Michigan must have been affected by the freedom of access to the open sea. Under conditions of unrestricted connection, the waters of inland basins into which no large rivers drained would have salinity approximating that of the open sea, but if connections were restricted and a moderately humid climate prevailed, the salinity would approximate that of the indraining streams. The drainage into the Huronian basins of Upper Michigan must have come from crystalline rocks or their weathered products and the waters would have contained only a few parts per millions of magnesium and calcium carbonate (Clarke, 1924, a, b). The waters would have been soft and the precipitation of lime carbonate by any agent or process would have been negligible. Such is the situation in existing lakes of northern Wisconsin and Upper Michigan which receive waters draining from crystalline rocks.

Deposition began in each of the major cycles of Huronian sedimentation in waters that were advancing over land areas and the initial deposits consisted of gravels and sands. The deposits became progressively younger in the directions in which the waters spread. Little is known of the relief of the

land surfaces that were submerged but the repeated occurrence of coarse gravels in Huronian sediments suggests that parts of the areas submerged were high and that islands existed in the Huronian waters. Island inliers of granite surrounded by coarse boulder conglomerates which grade into quartzite and slate are present in the W. 1/2 Sec. 22, T. 47 N., R. 25 W. and in the SW. 1/4 Sec. 9, T. 47 N., R. 25 W. Both of these areas are in the eastern part of the Marquette district and were described by Van Hise and Bayley (1897, pp. 220, 239-240, 275-277, 557). Marked relief of the Huronian surface is also indicated in the Sturgeon district where over 2,000 feet of coarse clastics are at the base of the Huronian section in that area.

Relation of the size, shape, and depth of the depositional area to the type of sediments and thickness of the lithic units.—The size and shape of the deposition area determined the magnitude of waves and velocity and extent of currents produced by waves. Wave dimensions and current intensities in turn determine the depths of the base levels of erosion and deposition. The quantity of sediments introduced, the capacity and competency of waves and currents, and depth of water are the chief factors that control distribution of sediments in standing bodies of water. Deep waters inhibit extensive distribution of gravels, sands, and even muds, whereas these sediments may attain wide distribution in shallow waters. Coarse sediments, in general, are deposited close to shores, colloidal and chemical sediments may come to rest over all parts of areas of deposition but make pure deposits only where clastic sediments do not mask them. Clastic sediments deposited in shallow waters are likely to be widely spread and the thickness of a depositional unit may be small, whereas in deep waters close to land relatively thick clastic units of limited extent may be formed from the same supply.

Relation of lithic units to time units.—Deltas are composed of three components: topset, foreset, and bottomset beds. Topset and foreset beds may contain coarse sediments associated with fine sediments, the bottomset component is usually composed of fine sediments only. The sediments of the three components which are deposited at the same time are stratigraphically equivalent regardless of differences in physical composition or differences in elevation with respect to sea

level. Thus the lithic units of the sediments of a delta transect the time units.

Sediments derived from erosion of shores are transported outward to deeper waters or are shifted by longshore currents into bay heads and thence, perhaps, outward along the axes of the bays. Coarsest sediments brought to a beach tend to remain, for waves moving toward shore generally have greater capacities and competencies than the returning waters. Coarse sediments are likely to be shifted seaward only during storms. With a stationary water level a wave-cut terrace is formed as a shore is eroded inland on which no permanent deposits are made except at the beach where gravels, sands, and even muds may be deposited. Deeper waters beyond the terrace receive deposits of muds, sands, and perhaps fine gravels. These sediments generally grade outward into the chemical and colloidal sediments which are deposited in the deeper waters.

With rise of water level, the muds, sands and fine gravels progressively invade the outer margin of the terrace where permanent deposition is possible as this part of the terrace comes below the base level of erosion. If rise of water level continues, chemical and colloidal sediments eventually follow the clastic sediments. Thus a section taken on the outer margin of the terrace will consist of clastics of some degree of coarseness at the base followed by finer clastics and these perhaps by chemical and colloidal sediments. With rising water level each lithic unit becomes younger as it progresses landward; it transgresses time units diagonally. The lithic units slope basinward at a steeper angle than the stratification.

If water level ceases to rise or remains stationary for a time, each lithic unit will progress basinward as rapidly as the bottom is built to the level at which each type of sediment may be deposited. The bedding will slope basinward at a steeper angle than the lithic units and sands will overlie silts, silts overlie clays, and clays overlie chemical and colloidal sediments.

If water level falls, each lithic unit will be extended basinward just as if water level were stationary. The lithic units will slope basinward at a flatter angle than the bedding and the basinward extensions will be younger than the landward parts even though they are deposited at lower elevations. Very complex sedimentary sequence may develop from still-

stands or lowering of water levels, which may be but temporary interruptions of a general rise of water level.

The sedimentary patterns described above are those developed on existing bottoms of standing water. Shallow water gravels grade into sands, sands grade into muds, and muds pass laterally into carbonate or other chemically or colloidally precipitated sediments. It seems certain that the chemically precipitated sediments in any Huronian section should similarly pass laterally into muds, these into sands, and sands into gravels.

Influence of relief and vegetation in source areas, and of distance of transportation, on the character of sediments.— The character and quantity of sediment transported to a basin of deposition is related to the vegetative cover of the source areas, the relief of source areas, and the distance through which the sediments are transported. Source areas of marked relief close to sites of deposition may furnish large volumes of coarse as well as fine materials. Similar areas long distances from sites of deposition are almost certain to furnish only fine sands, silts, and clays. The quantity transported, however, may be very great. This is well illustrated by the large quantity of fine sediments accumulating in the delta of the Mississippi River. Source areas of low relief do not supply significant quantities of fine or coarse clastics. The loads transported therefrom consist largely of colloids or dissolved materials. By the time an area has approached base level, the alkalies and alkaline earths have largely been removed as carbonates or bicarbonates to sites of deposition. Iron and silica tend to be removed more slowly, the former under acid, the latter under alkaline conditions, and they thus assume relative quantitative importance in streams draining thoroughly weathered terranes of low relief.

A well developed vegetative cover protects areas of marked and mature relief from mechanical erosion. Sands and coarse clastics are contributed to streams only where the vegetative covers are breached. Chemical weathering, however, is not decreased by vegetative cover and streams are likely to transport significant quantities of colloids and dissolved substances to sites of deposition. Little is known concerning the stage of development, abundance, or distribution of land vegetation during the Huronian, but the fact that many Huronian sedi-

ments were largely products of mature decomposition suggests that the land areas were protected by some form of vegetation. The large quantities of graphitic material in the Upper Huronian slates and the presence of algal structures in the Lower Huronian dolomitic limestones show that vegetation thrived in Huronian waters.

Relation between atmospheric composition and manner of transportation and deposition of iron compounds.—Iron may be transported in solution as ferrous bicarbonate, ferrous chloride, ferrous sulphate, or as colloidal ferric hydroxide or oxide. Transportation of ferrous bicarbonate requires an environment in which oxygen is deficient. The waters must also be charged with carbon dioxide. At the present time iron is not normally transported in surface waters in the ferrous state, but rather as colloidal ferric oxide or hydroxide. Formation of ferrous carbonate under existing conditions would require reduction of the ferric iron by associated organic matter after deposition.

Van Hise and Leith (1911, pp. 525-526) have suggested that at times during the Precambrian iron carbonate and silica may have been derived from volcanic or magmatic sources and contributed directly to sea water or through submarine springs. There is no doubt that deposition of iron carbonate from such sources would take place if the ferrous iron did not come in contact with oxygen of the atmosphere.

Behrend (1936, pp. 323-327) has observed and recorded data that are critical in the matter of deposition of iron carbonate from volcanic or magmatic sources. Submarine springs near the volcano Santorin in the Aegean Sea southeast of Greece have discharged large volumes of water containing iron carbonate in solution at various times since 1650 and the sea in the vicinity of the springs has been colored yellow, green, and red over considerable areas. The waters containing the iron carbonate change in color to a brownish yellow as the carbonate oxidizes. Iron oxide or hydroxide and not siderite is formed and deposited over the bottom. The accumulation of the extensive deposits of iron carbonate in the Huronian of the Lake Superior region would seem to require an atmosphere rich in carbon dioxide, burial of large quantities of organic matter with ferric iron if the atmosphere had anything like its present composition, or such rapid

deposition of iron carbonate that possibility of contact with oxygen was very limited.

Formation and preservation of stratification.—Although it seems to be essentially impossible for sediments to be deposited in water without development of some degree of stratification, muds and marls at present accumulating on the bottoms of all lakes studied in Wisconsin generally lack stratification. Doubtless it develops, but is destroyed shortly after deposition by mud-dwelling and mud-eating organisms which burrow through and work over the freshly deposited material for the contained organic matter. The perfect preservation of thin laminations in Huronian sedimentary rocks indicates that such organisms did not exist or that conditions in the sediments and overlying water were of such nature that these organisms could not live in the environment. The excellent preservation of laminations suggests that organisms of this habit had not yet evolved.

HURONIAN OF THE MARQUETTE DISTRICT

The Huronian rocks of the Marquette district were subdivided by Van Hise (1891, pp. 117-137), on the basis of a pronounced unconformity at the top of the chief iron-bearing formation, into the Upper and Lower Marquette series. Van Hise (1892, pp. 184-186) correlated the two subdivisions with the Upper and Lower Huronian of the original Huronian area of Ontario. Van Hise and Bayley (1897, p. 221) recognized that the Mesnard quartzite, Kona dolomite, and Wewe "slate" were confined to the eastern part of the Marquette district, and in order to explain the absence of these formations farther west they postulated that the sea in Lower Marquette time invaded the area from east and north and did not reach the western part of the area until Ajibic time. The stratigraphic succession of the Huronian in the Marquette district as outlined by Van Hise and Bayley (1897) is shown in figure 2.

Adams, Bell, Hayes, Leith, Miller, and Van Hise (1905, pp. 91-92), in a report of a special committee on nomenclature and correlation of the Precambrian formations of the United States and Canada, subdivided the Huronian sequence of Marquette district into three units—Lower, Middle and Upper Huronian. The base of the Middle Huronian was defined at

the bottom of the Ajibic quartzite where they state the "work of Professor Seaman has shown that there is a pronounced unconformity marked by strong basal conglomerates." Unfortunately a detailed description of this unconformity was never published by Professor Seaman.

The tripartite division of the Huronian was followed by Van Hise and Leith (1911) and they point out in tracing the Ajibic quartzite from east to west that it is successively in contact with the Wewe, Kona, and Mesnard units and finally with the Keewatin schists and Laurentian rocks. The Ajibic was assumed to cut diagonally across the beveled edges of formations ranging in age from the Wewe to Keewatin. Van Hise and Leith (1911, pp. 260-261) concluded that "These relations together with the presence of conglomerates at the base which bear debris from the Lower Huronian, show that the Lower Huronian was sufficiently indurated to yield fragments to the Ajibic before the deposition of that formation. The absence of the Lower Huronian in the western part of the district is doubtless largely if not wholly due to its removal by erosion between the time of the Wewe slate and the deposition of the Ajibic quartzite." Van Hise and Leith (1911, plate 19) presented a map by A. E. Seaman of the quartzite ridges near Teal Lake showing the exact location of the unconformity. The succession as accepted by Van Hise and Leith (1911) is given in table 1. They seem to have considered the several lithic units which they defined as sequential and seem to have given little consideration to the possibility that some of them might have been lateral equivalents of others.

The stratigraphy of the Upper Huronian of the Marquette district was revised by Swanson in 1929. He introduced the Greenwood formation to include a sequence of thin-bedded iron formation and magnetic shaly quartzites which lie conformably on the Goodrich formation. The opinion was expressed that the Greenwood formation had probably been included with the Bijiki in earlier mapping. Of far more stratigraphic value, however, is the fact that Swanson seems to have been the first to recognize that all the lithic units are of complex lithology and should be designated formations.

Leith, Lund, and Leith (1935) accepted Swanson's revision of the Upper Huronian sequence, but retained lithologic

TABLE 1

Revisions of the Stratigraphic Sequence of the Marquette District.

	Van Hise and Bayley, Mon. 28, 1897		Van Hise and Leith, Mon. 52, 1911		Leith, Lund and Leith, Prof. Paper 184, 1935		Tyler and Twenhofel, 1950	
					Acidic intrusives		Post Huronian—pre-Keweenawan granite	
Upper Marquette Series	Basic intrusives	Keweenawan	Basic intrusives Greenstone intrusives and extrusives	Michigamme slates	Basic intrusives		Basic intrusives	
	Michigamme formation	Upper Huronian	Michigamme slates		Upper slates Bijiki formation Lower slates	Upper Huronian	Michigamme formation: Upper slates Bijiki member Lower slate member	
	Clarksburgh formation		Clarksburgh volcanics		Clarksburgh volcanics		Clarksburgh formation	
	Ishpeming formation (Goodrich quartzite and Bijiki schist)		Bijiki schist Goodrich quartzite		Greenwood formation Goodrich quartzite		Greenwood formation Goodrich formation	
	Unconformity	Unconformity	Unconformity		Unconformity		Unconformity	
Lower Marquette Series	Negaunee formation	Middle Huronian	Negaunee formation		Negaunee formation	Lower Huronian	Negaunee formation	
	Siamo slate		Siamo slate		Siamo slate		Siamo formation	
	Ajibik quartzite		Ajibik quartzite		Ajibik quartzite		Ajibik formation	
	Wewe slate	Unconformity	Unconformity		Unconformity		Wewe formation	
			Wewe slate		Wewe slate			
	Kona dolomite	Lower Huronian	Kona dolomite		Kona dolomite		Kona formation	
	Mesnard quartzite		Mesnard quartzite		Mesnard quartzite		Mesnard formation	
	Unconformity	Unconformity	Unconformity		Unconformity		Uncomformity	
	Laurentian granite, syenite, peridotite, Palmer gneiss	Laurentian	Granite, syenite, peridotite, Palmer gneiss		Granite, syenite peridotite, Palmer gneiss		Granite, syenite, peridotite	
	Kitchi and Mona schist	Keewatin	Kitchi and Mona schist		Kitchi and Mona schist		Kitchi and Mona schist	

designations for the several units other than the iron-bearing formation (table 1). More recently W. A. Seaman (1943, pp. 11-17) advanced the idea that the unconformity reported by A. E. Seaman between the Lower and Middle Huronian is of major significance and was produced by a deformative movement which he termed the Champion Revolution. He assigned the Lower Huronian as defined by Van Hise and Leith (1911) to the Eozoic and subdivided it into two systems which he named the Kona and Mesnard. He also suggested the name of Iron for the Middle Huronian of Van Hise and Leith and the Michigamme for their Upper Huronian.

Study of the lithology, lithic sequence, and lateral facies changes of the Huronian section of the Marquette district has led the authors to suggest revisions of the stratigraphic section. The Kona dolomite, the Wewe argillite, and the Mesnard quartzite are not considered to be distinct time units but more or less lateral equivalents of a single time unit which is not given a name; the quartzite and the iron-bearing formation of the Palmer area (south of Negaunee, Michigan) are considered to be equivalents of the quartzite and iron-bearing units of the Siamo formation; and the evidence supporting the existence of an unconformity at the base of the Ajibic quartzite and between the Lower and Middle Huronian is critically examined and the view is reached that the unconformity does not exist. The environments of deposition of the Huronian sediments are interpreted according to the principles of sedimentation already outlined. The revised section according to the views of the authors is shown in table 1.

Mesnard Formation

Distribution.—The quartzite at the base of the Huronian section in the Marquette district was originally named the Mesnard by Van Hise and Leith. The view is later advanced that the Ajibic formation of the western part of the district may be continuous with the Mesnard of the eastern part. The discussion which follows applies only to the quartzite present at the base of the section over the eastern part of the westward pitching Marquette syncline.

The surface expression of the Mesnard quartzite is that of a horseshoe with the northern prong terminating east of Teal Lake near Negaunee and the southern prong about a mile west

of Goose Lake south of Negaunee. Most of the area of outcrop is covered with glacial drift, but locally the quartzite forms prominent hills. The type section is exposed upon the slopes and summit of Mount Mesnard south of Marquette, Michigan.

Lithology.—The lithology at the type locality and adjacent areas consists of a basal conglomerate succeeded by a clean quartzite followed upward by a shaly quartzite which passes into shale. East of Teal Lake the basal conglomerate is succeeded by interbedded shales and quartzites which pass upward into relatively clean white and pink quartzites. Van Hise and Bayley (1897, p. 224) divided the formation into four members: (1) a basal conglomerate, (2) slate and quartzite, (3) quartzite, and (4) slate. Later Van Hise and Leith (1911, p. 256) concluded there were but three members, a basal conglomerate, a central quartzite, and an upper slate. It is very doubtful if any division as simple as either of the above can be made, because the lithic sequence evidently varies greatly from place to place.

The basal conglomerate of the Mesnard was deposited on the older Keewatin greenstones and the Laurentian granites and gneisses. It has excellent exposures on the Keewatin greenstones north of Mud Lake in Sec. 29, T. 48 N., R. 25 W. and greenstone fragments which range to boulders 3 feet in diameter are abundant; pebbles of granite, gneiss, and vein quartz are present in subordinate quantity. East of Teal Lake the conglomerate also rests on the greenstones, but the larger particles are mostly well-rounded pebbles and cobbles of white vein quartz with subordinate greenstone particles. Excellent exposures of conglomerate lie on granite on the south side of the Marquette syncline west of Lake Peliesser in SW 1/4 Sec. 9 and SE 1/4 Sec. 8 T. 47 N., R. 25 W. This conglomerate is largely composed of angular to somewhat rounded pebbles, cobbles and large boulders of reddish granite, with occasional pebbles of white vein quartz. The conglomerate in these exposures is also interstratified with reddish arkose and white to pink quartzite and has the appearance of grading into the underlying granite.

The quartzite facies is composed of clean well-rounded to angular quartz sands which grade into shales, arkoses and conglomerates. Secondary enlargement and recrystallization of the original quartz grains have produced a rather vitreous

quartzite which locally is cross-laminated and contains ripple marks. In some areas the quartzites have a cherty appearance which apparently was produced by crushing of the original sand grains by dynamic forces followed by later recrystallization.

The slates range in color from black through various shades of gray to white and locally colors are red and brown. The black is due apparently to carbon and the white to an abundance of sericite. Because of relative incompetence the slates have been extensively sheared and commonly have well-developed flow cleavage.

Thickness and stratigraphic equivalents.—Van Hise and Bayley (1897, p. 231) recognized that since the Mesnard is a basal deposit of transgressing waters it must vary greatly in thickness due to irregularities of the surface on which it was deposited. They give the range in thickness from 150 to nearly 700 feet. However, the maximum thickness can not be determined at any one place and the total thickness would be a summation of the thickness of all beds from the beginning to end of deposition.

The Sturgeon quartzite of the Menominee district (locally a conglomerate over 2000 feet thick), the Saunders formations of the Iron River District, and the Sunday quartzite of the Penokee-Gogebic district are considered the general stratigraphic equivalents of what has been called the Mesnard quartzite. The correlations are made only in the sense that the quartzites all lie at the base of the Huronian section. They cannot be considered in each case to represent the same span of geologic time even though the composing sediments may have accumulated in the same basin of deposition. It is very probable that the deposition of clastics ended in some parts of the basin before it began in other parts.

Environment of deposition.—The various sediments that form the Mesnard formation do not seem to be fluvial deposits as the sands which form a large part of the unit are generally clean, fairly well sorted, and, in general, well stratified. Furthermore, such fluvial characteristics as cut-and-fill and the development of lenticular beds on a limited scale are not particularly apparent. The muds which formed the shales and slates interbedded with the quartzites also seem to have been well sorted and well stratified and do not have the features sug-

gesting stream deposition. There is no possibility that the environment was one of aeolian deposition.

The composition and sedimentary structures of the Mesnard formation and its stratigraphic equivalents seem to indicate that some of the sediments were deposited upon parts of the bottom of a shallow but perhaps extensive body of water. This water initially covered a small area, grew progressively larger, and ultimately covered much of Upper Michigan and northern Wisconsin. The great variation in thickness of the Mesnard suggests that the surface upon which it was deposited had at least 500 to 600 feet of relief. Furthermore, the deposition of 700 feet of gravel, sand, and mud indicates that the site of deposition must have been subsiding.

A strictly standing water environment of deposition can hardly be postulated for all of the sediments of the Mesnard and its stratigraphic equivalents, particularly for the more than 2000 feet of gravels and boulders that formed the Sturgeon quartzite and quartzite conglomerate of the Menominee district. These consist of beds of dirty quartzite and quartzite conglomerate with particles ranging from clay size to large cobbles. Attitudes are vertical. Such great thickness of coarse sediments could hardly have been deposited in a standing body of water, particularly as the sediments are not well sorted. These sediments probably were deposited on a piedmont delta or alluvial fan.

The sedimentary conditions that existed in the Mesnard basin of deposition should have been very similar to those existing in present standing water bodies. Gravels would be localized along beaches, but some beaches would be composed of sands, others largely mud. In places the beach and shallow water deposits would be interrupted by stream deposits in the form of deltas. In an ideal situation the gravels of beaches should grade outward toward deeper water into sand, the sand into mud, and finally the mud might grade into chemically or colloidally precipitated sediments. This ideal situation, of course, would be complicated by such factors as currents, irregularities of the bottom, and proximity to the mouths of rivers. The authors believe that the waters were saline, at least for a part of the time during which fine-grained and perhaps chemical equivalents of the Mesnard quartzite were deposited.

As the basin of deposition was widened and deepened, the

areas over which the basal sands and gravels were deposited were extended outward from the initial basin, and the sediments progressively were deposited at higher levels topographically as well as stratigraphically, therefore becoming progressively younger from the initial basin of deposition outward toward its borders. Thus, the middle and later parts of the Mesnard quartzite probably have time equivalence with the Kona dolomites and the Wewe argillites.

Kona Formation

Distribution.—The Kona formation is confined to the eastern part of the Marquette district and the type locality is in the hill country north of Goose Lake where numerous cliffs provide excellent exposures.

Lithology.—The Kona formation is largely a dolomitic limestone with numerous zones and beds of shale (slate in places) and sandstone or quartzite. Pink to white chert is present as lenses, small nodules and very small irregular bodies scattered through some of the limestones. Granular recrystallized facies of the chert resemble fine-grained quartzite. The limestone is generally very compact and ranges in texture from finely crystalline (often with lithographic texture) to very coarse with individual crystals to several millimeters in diameter. The range in color is from white to yellow and pink; the latter colors suggest the presence of siderite. Beds range in thickness from a few inches to about 4 feet. Thin lamination is common, but there is little or no separation along lamination planes. Laminae vary in thickness from a third to a tenth of an inch. Beds with lithographic texture are generally not laminated. The dome-shaped algal structures known as *Collenia kona*, rare in some localities, are abundant south of the city of Marquette where the strata are greatly deformed, in the Kona hills north of Goose Lake where the beds are nearly in horizontal position, on Lake Antoine where the attitude is vertical, and in the Bad River limestone east of Wakefield where the inclination is small.

Southwest of Goose Lake in Sections 22 and 23, T. 47 N., R. 26 W. the Kona grades to the west and south into calcareous shales. This was recognized by Van Hise and Bayley (1897, p. 255) for they state "Of the large exposures southeast of

Goose Lake, probably not more than one-third of the thickness is composed of reasonably pure dolomite, the remaining two-thirds being largely mechanical sediments." The shales in turn pass into quartzite and granite boulder conglomerate as the granite inliers in Section 22 are approached. On the north limb of the syncline east of Teal Lake in Section 32, T. 48 N., R. 26 W. the Kona likewise grades to the west into shales and quartzites[1]. The dolomite facies thus passes westward on both flanks of the Marquette syncline into clastic sediments.

Thickness and stratigraphic equivalents.—Van Hise and Leith (1911, p. 261) state that the Kona formation appears to range in thickness from a few feet to a maximum of 750. However, these authors did not recognize the extensive nature of the faulting in the Marquette district and it may be that the maximum thickness is not more than half the above figure.

The Kona formation of the Marquette district correlates so far as approximate position in the Huronian section is concerned with the Randville dolomite of the Sturgeon River, Felch Mountain and Iron Mountain districts, the dolomite in the Saunders formation of the Iron River district, and the Bad River dolomite of the Penokee-Gogebic district. The textures of these dolomites are very much alike except in the Felch Mountain district where the rock has been metamorphosed into a sugary marble. *Collenia kona* is present in all of them.

Transition between the Sunday quartzite and the Bad River dolomite and the presence of dolomite near the base of the Sunday quartzite are well displayed in the Penokee-Gogebic district in Sections 13, T. 47 N., R. 45 W. and 18, T. 47 N., R. 44 W. The strata are inclined 30 to 40 degrees to the north.

Descending section in SW 1/4 of NE 1/4, Section 13, T. 47 N., R. 45 W.

	Feet
Quartzite, gray	1.0
Dolomite, thinly laminated, purple to pinkish-gray, numerous chert beds	6.0
Quartzite, gray	.5
Dolomite, thinly laminated, pale purple to purplish gray	4.5
Quartzite, vitreous, fine-grained, gray	1.3

(1) The authors were unable to detect any dolomite in the rocks mapped by A. E. Seaman as dolomitic slate and quartzite (Van Hise and Leith, 1911, plate 19).

Quartzite, shaly, thinly-bedded with a lens (0-3 inches thick) of grayish-red dolomite	1.0
Quartzite, light purple, poorly sorted grains	1.0
Quartzite, gray with pebbles of chlorite schist	.8
Quartzite, pale greenish-gray, fine-grained with thin shale partings	1.0
Quartzite, purplish-red, shaly	.1
Quartzite, dark purplish-gray, with ferrugineous argillite pebbles	.5
Quartzite, pinkish-gray, fine-grained, cross-laminated with pebbles and cobbles (up to 2.5 inches) of fine-grained quartzite, argillite, chert and dolomite	16.0
Quartzite, pinkish-gray, vitreous, cross-laminated with foresets inclined northerly, red shale pebbles	36.0

The basal zone of the above section is not far above the base of the Huronian sequence. Lenses of dolomite contained in quartzite and interbedding of dolomite and quartzite indicate that both sands and chemical sediments were being deposited more or less simultaneously.

Descending section in SW 1/4 of SE 1/4 Section 18 T. 47 N., R. 44 W.

	Feet
Dolomite, thinly laminated, fine-grained with semi-conchoidal fracture, pinkish-gray. Beds up to 4 feet thick. Lenses and bands of chert up to 4 inches thick, *Collenia kona* present throughout	100.0
Dolomite, pinkish-gray, interbedded with greenish shale	1.0
Quartzite, vitreous, evenly-laminated, gray	2.0
Shale, thinly-laminated, greenish-gray	1.0
Quartzite, thinly-bedded, locally cross-laminated, with greenish-gray shale partings up to 1 foot thick	30.0
Dolomite, pinkish-gray	3.0
Quartzite, grayish-pink, rather coarse grained and somewhat arkosic	2.0
Conglomerate, white quartz and green schist pebbles up to 2 inches in diameter	0.1-10
Greenstone schist (Keewatin)	10.0

This section shows that interbedded shale, dolomitic limestone and quartzite are present about 2 feet above the base of the Huronian and at this place the Sunday quartzite and the Bad River dolomite seem to have been deposited during essentially the same span of geologic time.

Environment of deposition.—The several limestones of the Huronian section of Upper Michigan accumulated in areas of a depositional basin that were at times free from the entrance of sand and mud. The interbedded quartzites indicate that clastic sediments from time to time were introduced over the areas where chemical sediments were in process of deposition. These areas either lay off shore from the areas of deposition of sands and muds or close to shore in areas where clastics

were ordinarily not available. Cherts that represent deposition of silica with the carbonates suggest that land waters were mingling with marine.

The abundance of calcium and magnesium carbonates in the Lower Huronian of Upper Michigan strongly suggests that the basins of deposition of this area had unrestricted connection with the open sea so that there was mutual circulation. Under such conditions the waters could readily have been high in calcium and magnesium carbonates and deposits could have been made by lime-secreting algae. The Kona may represent an algal reef limestone similar to those formed in the Pennsylvanian seas. Waters derived from a crystalline land area accumulating in a basin with restricted connection with the open seas would be likely to be low in calcium and magnesium carbonate, as are the present day soft water lakes of Upper Michigan.

That land bordering the areas of deposition had sufficient local relief to move clastics to the sites of deposition is shown by the quartzites and argillites interbedded with the dolomite. The carbonates that were primarily precipitated may have been aragonite, calcite, or dolomite. The compactness of horizontal lithographic texture of the dolomites suggests that, if these were not primary precipitates, the original carbonates were altered to dolomites prior to lithification. The frequency of well-developed laminae may have a bearing on this problem. Perhaps carbonate mud was from time to time thrown into suspension and acquired the needed magnesium from the suspending waters. Differential settling of the carbonate mud, colloidal silica, and clay may have produced the laminae, which were preserved because mud-eating and mud-boring animals were absent.

The dome-shaped structures of *Collenia kona* now consist of alternate laminae of dolomite and chert or entirely of chert. Present composition is considered to be secondary. The laminae are concave downward and are stacked much like a sequence of inverted bowls with the smallest bowl at the base and progressively larger bowls upward. Single colonies up to 6 feet high and 3 feet wide at the top are not uncommon. Most students have thought these structures to be of algal origin because of close resemblance to similar structures made by living algae. W. A. Seaman (1943, p. 13) states that the

structures are related to deformation, a view that the authors believe does not merit serious consideration.

Wewe Formation[2]

Distribution.—The type locality of the Wewe as defined by Van Hise and Bayley (1897, p. 256) is in the Wewe Hills south of Goose Lake. Van Hise and Leith (1911, plate 17) assigned the slates of the type locality to the Mesnard and restricted the term to the argillites and graywackes which lie between the Kona dolomite and the Ajibic quartzites east of Teal Lake and north of Goose Lake. The field evidence shows that the Wewe rocks of the original type locality are the lateral equivalents of the Mesnard quartzite and the Kona dolomite and it is believed that a like relation applies to much of the Wewe as restricted by Van Hise and Leith.

Lithology.—The Wewe is well exposed on Carp River in the SE 1/4 Sec. 6, T. 47 N., R. 28 W. The rocks range from gray to black and from brown to red argillites. Parts are massively bedded, others consist of thin alternating laminae of light gray and darker gray to black shale or argillite. There is little metamorphism. The number of laminae per inch ranges from 10 to 16 with the lighter colored laminae in some cases thicker than the dark colored. Laminae are commonly very regular but some show local crumpling.

Thickness and stratigraphic equivalents.—The thickness of this formation is not known, but it seems to vary considerably. Van Hise and Bayley (1897, p. 271) state that the slate in Sec. 22, T. 47 N., R. 26 W. is 1050 feet thick, and in Sections 13 and 24 it is barely 100 feet thick. These strata were later included by Van Hise and Leith in the Mesnard (1911, plate 17), but it is interesting to note that they (1911, p. 259) retained the figures of 1050 and 100 respectively for the thickness of the Wewe. The thickness of the exposure on Carp River is only about 100 feet.

The Wewe formation is limited to the eastern part of the Marquette district and it has no recognized lithologic equivalent in the Huronian of other districts in Michigan.

[2] The Wewe is designated a formation largely on the basis of courtesy. It is not considered a mappable unit.

Environment of deposition.—The uniformity of bedding and the fine-grained character of the sediments indicates deposition below wave base in a standing body of water. Evidence of strong current or wave action seems to be entirely wanting. The excellent laminae suggest periodic fluctuation in the kind of sediment deposited. The dark bands probably accumulated when organic matter flourished in the overlying waters and a pair of light and dark laminae may represent an annual deposit. If this is the case, the rate of deposition of the Wewe argillites would have been about one inch of sediment in 5 to 8 years. The argillites grade on the one hand into the sands of the Mesnard quartzites and in the basinward direction into the Kona calcareous sediments and thus the three formations to some degree are time equivalents. Relations of this kind are well known in later parts of the geologic column and have been described by McKee (1945, pp. 3-168) in the Cambrian of the Grand Canyon region and by Stephenson and Monroe (1938, pp. 1639-1657) in the Upper Cretaceous of Alabama and Mississippi.

Ajibic Formation

Distribution.—The Ajibic formation as defined by Van Hise and Bayley (1897, pp. 282-513) rests directly on the older gneisses and granites over the western part of the Marquette district and over the older Huronian strata in the eastern part. The type locality is in the Ajibic Hills a short distance northeast of Palmer where there are excellent exposures.

Lithology.—In the western distribution the Ajibic consists of a basal conglomerate containing pebbles and cobbles of the older granites, gneisses, and schists overlain by clean vitreous quartzites, argillaceous quartzites, and quartzites interbedded with argillites. In the eastern distribution in the Ajibic Hills the formation consists of massively to thinly bedded clean quartzites, with lenses of conglomerate at several horizons; it grades downward and laterally into the argillites and quartzites of the underlying Huronian. In the Palmer area the formation has a boulder conglomerate at the base and this rests on the Palmer gneiss. The lithology of the Ajibic is essentially similar to that of the Mesnard.

In the NW 1/4 of Section 27, and also in the NE 1/4 of Section 28, T. 47 N., R. 27 W. amaygdaloidal flows are interlayered with ferruginous mica-slates. Van Hise and Bayley (1897, p. 312) correlated these rocks with the Ajibic, but as the flows are in a stratigraphic horizon that is only a short distance below the base of the Negaunee formation it is probable that the flows represent the volcanism that is associated with the Goose Lake member of the Siamo formation.

Thickness and stratigraphic equivalents.—According to Van Hise and Leith (1911, p. 261) the formation ranges in thickness from a few feet to a maximum of perhaps 750 feet. The only good section shows a little more than 100 feet.

The Felch schist of the Felch Mountain district and the Palms formation of the Penokee-Gogebic district are usually considered the lithic equivalents of both the Ajibic and Siamo formations of the Marquette district.

Relation to adjacent formations.—The Ajibic formation grades upward into the overlying Siamo formation without any indication of a stratigraphic break and it is probable that some of the sediments assigned to the Siamo were deposited at the same time as the Ajibic sands.

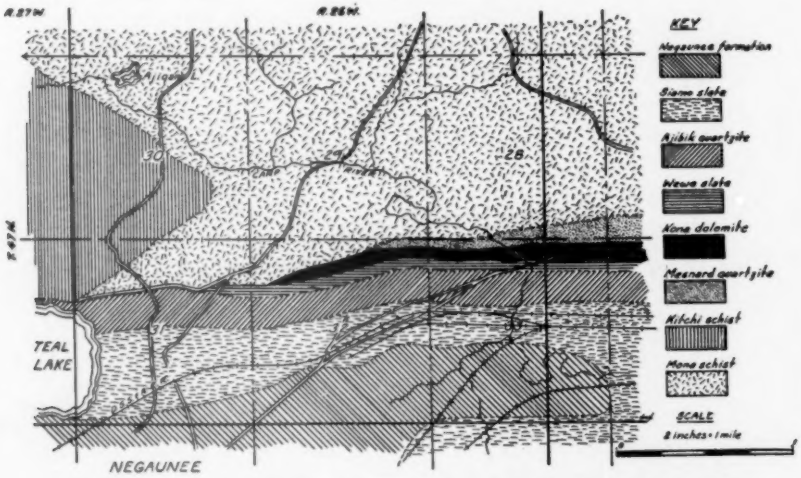

Fig. 2. Map of the area east of Teal Lake showing transgressive overlap to the westward of the Mesnard, Kona and Wewe. Interpretation by Van Hise and Bayley (1897).

The conglomerate at the base of the Ajibic in the western part of the Marquette district has unconformable contact with the underlying Keewatin granites, gneisses and schists. In the Teal Lake area in the central part of the Marquette district the Mesnard formation underlies the Ajibic and is considered conformable thereto by the authors.

Van Hise and Bayley (1897, p. 221) found no evidence of an unconformity between the Ajibic and the underlying formations of the Huronian. They state "At the beginning of Lower Marquette time the transgression of the ocean was from the east and north, and as a consequence the inferior formations of the Lower Marquette series appear only in the northeastern part of the district." They concluded that the western part of the Marquette district was above water and undergoing erosion at this time (fig. 2).

A. E. Seaman identified a rock as a conglomerate in the quartzite ridges east of Teal Lake and considered this conglomerate as the basal zone of the Ajibic. This identification led Van Hise and Leith (1911, p. 260) to conclude that the apparent transgressive overlap of the Ajibic in this area was due to unconformable relations of the Ajibic with the Wewe, Kona, and Mesnard formations. The acceptance of the un-

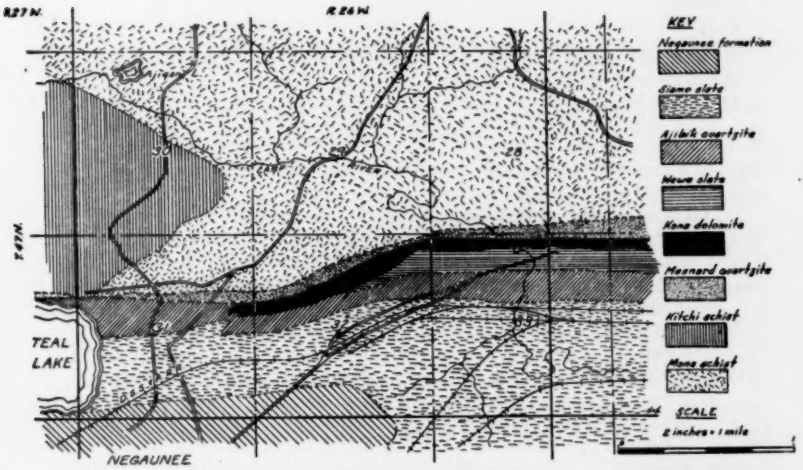

Fig. 3. Map of the area east of Teal Lake showing the Ajibik resting unconformably upon the Mesnard, Kona and Wewe. Interpretation by Van Hise and Leith (1911).

conformity led Van Hise and Leith to subdivide the Lower Marquette series of Van Hise and Bayley into the Lower and Middle Huronian and to assume that the Lower Huronian had probably been removed by erosion over the western part of the Marquette district although Seaman's map (Van Hise and Leith, plate 19) shows that the Kona and Wewe formations pass westward into quartzites (fig. 3).

W. A. Seaman (1943, pp. 13-14) greatly emphasized the magnitude of the postulated unconformity at the base of the Ajibic and he stated there was extensive deformation and erosion preceding Ajibic deposition. The deformation was termed the Champion Revolution. He also removed the Wewe, Kona, and Mesnard formation from the Huronian and assigned them to the Eozoic. The authors do not accept these views.

As a proper understanding of the relations of the Ajibic formation to the Mesnard is essential to the interpretation of the geology of the Marquette district, the present authors made detailed study of the extensive exposures east of Teal Lake where the unconformity at the base of the Ajibic was mapped by A. E. Seaman (Sec. 31, T. 48 N., R. 26 W.). A cut in the quartzite ridge was made in the construction of U. S. Highway 41 which exposed the following descending section.

Zone	*Feet*
16. Quartzite, light pinkish-gray with red spots, fine angular grains	30
15. Quartzite, light pinkish-gray, fine angular grains	20
14. Quartzite and slate interbedded. Quartzite fine-grained, gray. A boudinage structure, which resembles a conglomerate, has been developed in some of the quartzite beds interbedded with the slate. This zone is concealed in the cut but exposed eastward in the ridge	50
13. Quartzite, light pinkish-gray, vitreous, beds variable in thickness and regularity. Grains angular to subangular. The top of this zone contains thin shale lenses and the boudinage structures which A. E. Seaman identified as a boulder conglomerate. It is at the base of this supposed conglomerate zone that Van Hise and Leith place the base of the Ajibic and the top of the Mesnard	61
12. Quartzite, thick-bedded, light pinkish-gray, sugary to vitreous, grains angular. Irregular current ripple marks developed at base of the cliff. Counterparts of wave ripple marks are well developed on the bottom of a bed 7 feet above the base of this zone	12
11. Quartzite, thick-bedded, light pinkish-gray, sugary to vitreous, grains angular. Bed at top has current ripple marks indicating northerly moving currents	47

10. Quartzite, thick-bedded, light pinkish-gray, compact, vitreous, grains angular. One bed has cross-lamination with northerly inclination and foresets of 2 feet with vertical component of 6 inches 3
9. Quartzite, evenly bedded, light pinkish-gray, sugary to vitreous, grains angular. Beds 4 inches to 1 foot thick. Current ripple marks present at two horizons 21
8. Slate, white, composed of quartz and sericite 6.5
7. Slate breccia, black, in quartzite matrix. Grades into black slate. Breccia probably dynamic. Fragments angular 5
6. Slate, black 1
5. Quartzite, gray, with one bed of black slate breccia in quartzite matrix. Breccia probably dynamic 1.5
4. Slate, black and white interbedded with gray quartzite 10
3. Quartzite, gray, vitreous, with some interbedded black slate 10
2. Concealed, estimated thickness 50
1. Conglomerate with abundant rounded pebbles and cobbles of white vein quartz and a few subangular fragments of green schist 3

Zone 13 consists of interbedded quartzite and slate which is extensively sheared and contains typical boudinage structure of rounded to subangular quartzite fragments in a slaty matrix. This is the material that A. E. Seaman mapped as conglomerate and considered evidence of an unconformity. However, similar breccia and boudinage structure are present in this quartzite ridge at many horizons and some of them have much greater resemblance to a conglomerate than the one mapped by Seaman. Vein quartz pebbles, cobbles and boulders which are so generally characteristic of the Huronian conglomerates of the Marquette district are totally absent in the rock designated conglomerate. The only material is quartzite like that of the associated beds. If this were a true conglomerate it would seem that vein quartz particles should be present.

An eighth of a mile east of U. S. Highway 41 on the south side of the quartzite ridge Seaman's map shows the supposed unconformity to be offset to the south by a fault. A rock which resembles a quartz pebble conglomerate is exposed on the eastern block for some distance along the strike of the supposed unconformity. The "pebbles" have spheroidal shapes, project slightly above the surface of the enclosing rock, and are whiter than the matrix. Borders are quite irregular, grade into the surrounding rock, and are not sharply defined. They are better cemented parts of the rock in which they are present.

It is possible that these pseudo pebbles were mistaken for vein quartz pebbles.

The strata above and below the unconformity postulated by A. E. Seaman appear to be strictly concordant and there is no indication of divergence of dip or strike to suggest an unconformity. The authors have no fundamental objection to placing the base of the Ajibic at the horizon indicated by A. E. Seaman, but evidence of an unconformity is considered to be totally wanting and they see no reason for making a division at the horizon selected by A. E. Seaman. It seems that Van Hise and Leith accepted the unconformity identified by A. E. Seaman largely on the basis that it explained the disappearance of the Wewe and Kona formations over the western part of the district for they state (1911, p. 260) that "In the eastern part of the district the Ajibic quartzite rests upon the Wewe slate, somewhat farther west on the Kona dolomite, and still farther west on the Mesnard quartzite—that is, it cuts diagonally across these three formations."

The present study shows that both the Kona dolomite and the Wewe argillites grade laterally into quartzites and conglomerates of the Mesnard formation. A similar change in lithology in the area east of Teal Lake causes the Kona and Wewe to disappear westward into quartzite and this is also shown in the map by A. E. Seaman. If this interpretation is correct, there is no unconformity at the base of the Ajibic in the Teal Lake area and the quartzite west of Teal Lake, which has been mapped as Ajibic, *may represent Mesnard, Kona, Wewe, and Ajibic time.*

In the Palmer area the Ajibic formation rests unconformably on the Palmer gneiss. Van Hise and Bayley (1897, pp. 211-218) considered this gneiss to consist of extensively sheared and silicified Archean granite. Lamey, however, (1935, pp. 1137-1161) has mapped materials in the Palmer gneiss belt south and east of Palmer which he identified as dolomite, quartzite, and graywacke and correlates these sediments on the basis of lithologic similarity with the Kona dolomite and Mesnard quartzite. If this correlation is correct, an unconformity exists between the Mesnard and Ajibic formations in the Palmer area. Its extent is not known. To assume that it is extensive presupposes a greater knowledge of Huronian paleogeography than exists at the present time.

Environment of deposition.—The excellent sorting, well-developed stratification, and presence of current and wave ripple marks suggest that the sediments that form the Ajibic formation were deposited in a standing body of water. This may have been in connection with a delta. The basal conglomerate in the western part of the district may represent a shore environment. The Ajibic sediments certainly become progressively younger in the directions in which deposition expanded.

(*To be continued*)

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THE D'ENTRECASTEAUX REEF GROUP

FRED R. HAEBERLE

ABSTRACT. The D'Entrecasteaux Reef Group, located east of Australia and north of New Zealand, is composed of two atolls, two platform reefs and two outer or barrier reefs. The structural arrangement of the group seems to indicate that the reefs grew on a submerged continuation of the New Caledonian block. Depths immediately surrounding the atolls and the general arrangement of the reefs, atolls in the center surrounded by barrier reefs, lead to the conclusion that the reefs may be more easily explained as growing on an antecedent platform, than by the submergence of a volcanic peak or the raising and lowering of sea level.

INTRODUCTION

THE D'Entrecasteaux Reef Group lies 800 miles east of Australia and 1300 miles north of New Zealand, at 18 degrees South latitude and 163 East longitude. It is approximately 80 miles northwest of the island of New Caledonia. A pass known as the Grand Passage, about 40 miles in width, separates the reef group from the northwestern end of the barrier reef surrounding New Caledonia. Although the passage is considerably shallower than the water on either side, where depths of over 11,000 feet have been measured, it is deep enough, 20-250 feet, to permit the safe passage of shipping.

The region was first visited by Captain Cook in 1744 who landed on the northeast coast of New Caledonia. Cook passed either through the Grand Passage or north of the D'Entrecasteaux Reefs, as he made no mention of them. They were first found in 1793, by the French explorer D'Entrecasteaux, after whom they were named.

Some confusion seems to exist regarding the correct spelling of the name D'Entrecasteaux. While the British and United States Hydrographic Offices prefer to use D'Entrecasteux, the name is French and the spelling in this paper will follow the French usage. The reef group under discussion here should be distinguished from another group of the same name lying northeast of the east end of New Guinea.

THE REEF GROUP

The Reef Group is composed of two major atolls, Huon and Fabre, two outer or barrier reefs, and two small platform reefs. The group trends in a northwest-southeast direction.

It is about 50 miles long and 35 miles wide, with an overall area of 1750 square miles. Figure 1 shows the relationships of the various reefs and was adapted from a series of charts of South Pacific islands.

From soundings taken around the group, it appears that this entire reef group grew on the top of an irregularly shaped, relatively flat-topped platform. Figure 1 shows the 50 fathom contour drawn around the group and outlines the maximum limits of this platform.

The northernmost of the two atolls, Huon, is somewhat horseshoe-shaped, with the horns extending in a northwesterly direction. As the prevailing winds in this region are from the southeast and east-southeast, this direction of growth is not surprising (Umbgrove, 1929). Figure 1 shows several depth readings of over 4,000 feet immediately west and southwest of the atoll, indicating that the platform upon which reef growth took place was not as fully developed on the western side as it was on the eastern. As a result, the southwestern horn could not grow as extensively as the one on the opposite side.

There is one well developed permanent island, Huon, in this atoll, although the greater part of the reef is above sea level at high tide. Passes through the reef exist only on the northwest, or lee side. The atoll is approximately 22 miles long and 10 miles wide. Measured depths inside the lagoon are as great as 33 feet, and possibly greater depths exist. Coral heads and patches are numerous within the lagoon and the bottom surface is very irregular. The reef is from 1,000 to 2,000 feet wide, reaching its maximum width along the southern edge.

The southern atoll, Fabre, is better developed than Huon Atoll, as the horns of its horseshoe have almost completely closed. One pass is present on the northwest side and two smaller passes are found on the southwest side of the atoll. Two islands, Fabre and Le Leixour, are located on the north side of the atoll, and one, Surprise, is located on the southwest side. The atoll is approximately 26 miles wide and 21 miles long. In the southwestern part of the lagoon, depths as great as 60 feet have been measured. Coral heads and reef patches are numerous inside the lagoon and the bottom surface is very irregular.

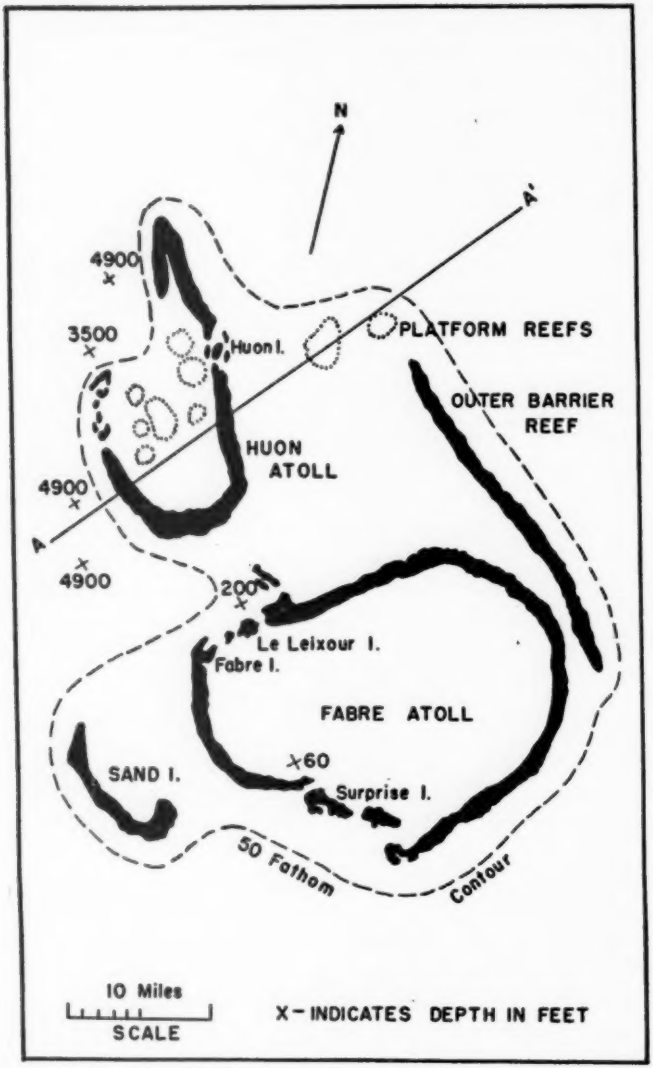

Fig. 1. Outline map of the D'Entrecasteaux Reef Group. Reefs shown in black. Reefs at or just below sea level indicated by dotted lines. Dashed line shows position of 50 fathom (300 feet) contour. Modified from U. S. Naval Hydrographic Office Chart.

No sounding have been taken in the pass between the two atolls which are 7 miles apart. One depth of 200 feet has been recorded immediately northwest of Le Leixour Island.

A ribbon reef northeast of the two atolls is 25 miles long and varies in width from 1300 to 2000 feet. A few breaks or passes exist in the reef, but none are large enough to permit the passage of even coastal shipping. At its nearest point the reef is 15 miles from Huon Atoll and 1½ miles from Fabre Atoll. The water depth between the atolls and the small barrier reef rarely exceeds 60 feet.

Southwest of Fabre Atoll is a small reef known as Sand Island, approximately 1800 feet wide and 10 miles long. The tips of the reef curve back toward Fabre Atoll much as would the ends of a sand spit or growing atoll. The direction of this bending would suggest that the prevailing winds come from the southwest, but as stated above they are from the southeast. Soundings taken offshore from Sand Island indicate that the reef grew where the offshore side of the platform begins to slope sharply to the deeper parts of the ocean. The inward bending of the northern part of this reef reflects an indentation in the platform along which the reef was growing. The inward bending of the southern tip is caused in part by the shape of the platform and in part by the direction of the prevailing winds, which tend to curve this part of the reef toward the north. Thus, a horn is being formed similar to those found on the atolls. Such curving is absent on the other barrier reef as it apparently grew along the straight edge of the platform, parallel to the direction of the prevailing winds. Whereas other reefs in this group have much the same composition as reefs elsewhere, Sand Island is characterized by a medium-to-coarse-grained calcareous mud, composed of tremendous quantities of broken and groundup fragments of shells, corals and algae.

Between Huon Atoll and the northern end of the northeast barrier reef are two platform (Fairbridge, 1950, p. 345) or table (Ladd, 1950, p. 204) reefs. The smaller is almost circular in shape and is 1½ miles in diameter; the larger reef is ovate and is 3 miles long and 1½ miles wide. Considerable portions of these reefs are above high tide level, and the remaining parts may easily be located by breakers. No permanent islands are present on these reefs.

Sand Island and the northeasterly reef may be called fringing reefs, but they are also small barrier reefs. The seaward side of each drops off sharply to the abyssal depths of the ocean, whereas the inner slope is comparatively gentle. Thus these reefs apparently grew along the edge of the platform. Figure 2 shows the steep slope on the flanks of the reef group.

ORIGIN OF THE D'ENTRECASTEAUX REEF GROUP

For many years there has been spirited controversy concerning the origin of coral reefs. Advocates of each theory claim that it is the only acceptable one. The author believes that different origins may be necessary to account for the different kinds of coral reefs known today. Perhaps one set of conditions influenced reefs in one area, but was not important in another area. The reefs in this group are believed to have been formed on the top of a sea mountain or upfaulted block of the sea floor, in accordance with the antecedent platform theory (Shepard, 1948, p. 251).

Physical evidence, in the form of outcrops, to prove a direct relationship between the main island of New Caledonia and the platform for the D'Entrecasteaux Reef Group is absent. However, circumstantial evidence is strong. New Caledonia rises abruptly from the great depths of the ocean as does the seamount. The northwest-southeast alignment of New

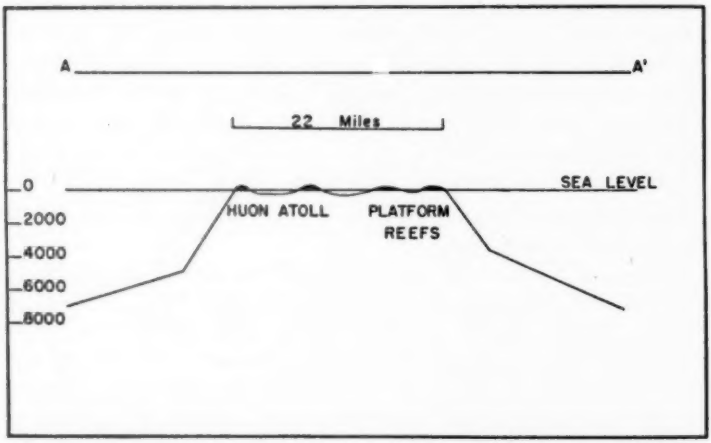

Fig. 2. Cross section along line A-A' through Huon Atoll and the Platform Reefs. Vertical scale greatly exaggerated.

Caledonia is continued in the platform and the reefs. In addition, the two areas are separated only by a narrow, shallow passage. Since the island of New Caledonia is composed mainly of serpentine (Compton, 1917, p. 84), it is reasonable to suppose that the base for the D'Entrecasteaux Reefs is composed of the same material.

The origin of the seamount upon which the reefs grew may be explained in at least two ways. It is believed that the southern end of New Caledonia has been uplifted while the northern end was depressed in fairly recent times (Compton, 1917, p. 100). The Grand Passage would then represent part of the downwarped portion. The base for the reefs is either a high area that has remained comparatively high despite the regional downwarping, or an area that has been upfaulted while the rest was being downwarped. If the area was downwarped, the reefs grew upward as the block slowly sank. If the area was upfaulted, it was raised enough to enter the zone where growth of reef organisms could take place.

The two outer reefs with the two atolls located between them make it difficult to apply Darwin's ideas to this area. Several separate periods of sinking of the central island would be necessary to explain this arrangement. The relatively flat-topped platform upon which the reefs grew does not appear to represent the type of structure presented by a sinking central island.

Assuming the type of rock in the seamount is the same as that of New Caledonia, an area of 1750 square miles of serpentine would have had to be eroded in an extremely short time, geologically, to explain these reefs by the glacial-control theory.

SUMMARY

The structural alignment of the reefs and the seamount, and the size of the area involved, favor the concept that the D'Entrecasteaux Reef Group grew on a large submerged extension of the New Caledonia block. The break in slope between the platform and the steeply sloping flanks of the seamount is the site of elongated ribbon or barrier reefs, except in the northwest corner where the platform drops off too abruptly. The reefs grew upon the relatively flat top of a seamount as soon as the proper environmental conditions

were present. The shape of the atolls is influenced primarily by the direction of the prevailing winds, whereas the shape of the barrier reefs is influenced by the direction of the winds and the outline of the platform. It would seem that this group is an excellent example of reef growth on the kind of antecedent platform assumed in the theory of Hoffmeister and Ladd (1944), which may not be as satisfactory a theory for some other reefs.

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MASS-WASTING ON TABLE MOUNTAIN, FREMONT COUNTY, COLORADO

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ABSTRACT. Table Mountain is a mesa-like feature in the foothills belt along the eastern margin of the Colorado Front Range. The resistant capping consists of the Cretaceous Dakota sandstone and the immediately underlying Fuson shale and Lakota sandstone. The latter rests on the thick shales of the upper Morrison formation. This sequence of rocks and the stream dissection which has largely isolated the mountain have combined to produce conditions favorable for extensive mass-wasting. Three types of mass-wasting—talus slumping, block slumping, and rock creep—are described and the factors influencing each are discussed.

INTRODUCTION

THIS paper describes an excellent example of mass-wasting which is readily accessible for field examination. Table Mountain is located in eastern Fremont County, Colorado, 23.5 miles southwest of Colorado Springs on Colorado Highway 115. Most of the area described here is shown on the old Colorado Springs Quadrangle map and the general geology is shown on the geologic map in the Colorado Springs Folio, No. 203. New 7½ minute quadrangle maps covering the area of the old 30 minute quadrangle have recently been issued by the United States Geological Survey. Table Mountain is located on the Mount Pittsburg sheet which has a scale of 1:25000 and a contour interval of 40 feet. Anyone wishing to visit the area should secure this latest map.

GEOLOGY OF TABLE MOUNTAIN

The top of Table Mountain, as the name suggests, is relatively flat. The area of the mountain top is about one and one-half square miles. This table-land slopes south and southwestward about 500 feet per mile. The mountain reaches its greatest height at the north end where it rises abruptly above Banta Gulch with a total relief of more than 1,000 feet. To the south the flatness of the mountain is broken by numerous canyons but the flat interfluves extend to the valley of Beaver Creek which is about 5 miles from the crest of the mountain. The eastern and western bedrock slopes are almost completely buried by debris produced by mass-wasting. The principal area of mass-wasting is located southeast of the mountain

crest and extends for more than a mile down a slope of about 500 feet per mile into upper Salt Canyon.

The top of the mountain is essentially a dip slope of the Cretaceous Dakota sandstone. The Dakota has a maximum thickness on the mountain of only 40 feet, as the upper portion of the formation has been removed. The sandstone is typical of the Dakota except that it appears more quartzitic on Table Mountain than elsewhere in the immediate area. The Dakota is underlain by the Fuson shale, about 50 feet thick although nowhere well exposed on the mountain. The Fuson is mainly black shale with local sandstone or sandy lenses and 3- to 6-inch beds of bentonite in the upper part. The Lakota sandstone underlies the Fuson and is also about 50 feet thick but poorly exposed on Table Mountain. The Lakota here is characterized by conglomeratic beds with grains up to an inch or more in diameter. It is much less quartzitic than the Dakota above. The Lakota is underlain by the Jurassic, Morrison formation, the upper portion of which consists mainly of shales. The lower Morrison contains massive sandstones, lithographic limestones and gypsum as well as shale. Permo-Carboniferous Red Beds and older Paleozoic rocks underlie the Morrison but they are of no consequence in the mass-wasting described here.

Structurally, Table Mountain is located at the junction of the southwest limb of the Red Creek anticline and the typical hogback ridge which extends northeastward from the Canon City embayment. The area of the mountain includes a number of minor sags or warpings of the beds but in general its structure is that of a southeastward dipping monocline with dips nowhere exceeding 10 degrees.

Topographically Table Mountain could be generally defined as a tilted table-land bordered on the north by a very steep scarp and on the east and west by slopes of varying degrees of steepness, and losing its identity to the south where the Dakota sandstone dips beneath the overlying Upper Cretaceous beds in Beaver Creek valley. In detail, however, the topography is a maze of deep canyons and gullies which by headward extension have carved the borders of Table Mountain into a network of steep slopes separated by sharp-crested to relatively broad and flat interfluves. The canyons are particularly well developed in the southern section of the moun-

tain where they have divided the mountain top into a number of long relatively flat-topped spurs. Landforms controlled wholly or in part by mass-wasting are characteristic of the topography along the steep western and northern slopes of the mountain and on the eastern side.

MASS-WASTING

Mass-wasting occurs in a number of distinct areas which are shown in figure 1. Areas 1 through 5 have many characteristics in common but area 6 is essentially unique.

We recognize three distinct types of wasting which may for convenience be designated: a) talus slump, b) block slump, and c) rock creek. The characteristics of these types may be described with reference to specific areas on the mountain.

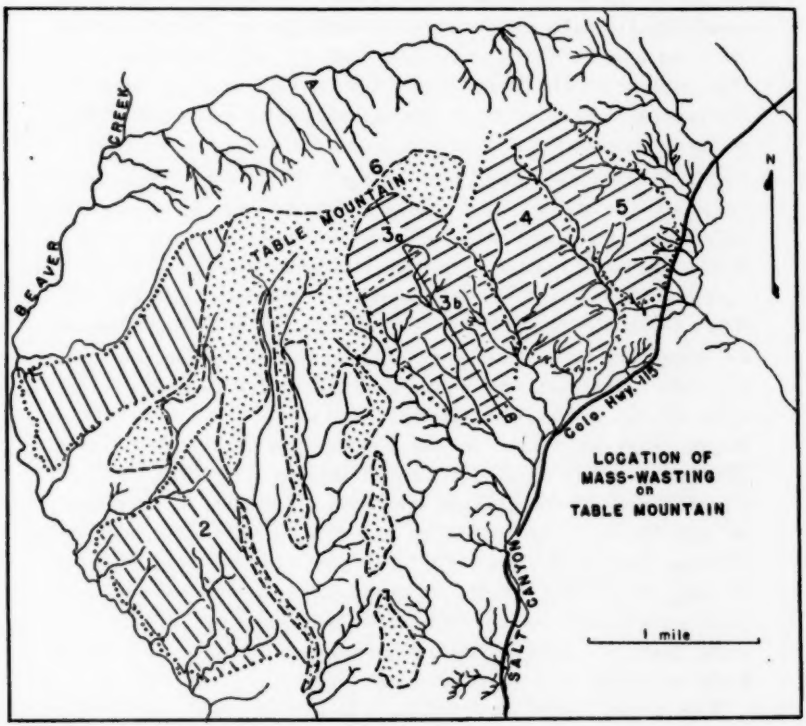

Fig. 1. Block slumping is well developed in areas 1, 2 and 3a; rock creep in areas 3b, 4 and 5; talus slumping in area 6.

Talus slumping.—This type of wasting is best represented, and is the only type of wasting, in area 6. Here the mountain face slopes steeply to the north whereas the capping sandstone dips gently southeast.

The mountain is in effect a high cuesta facing north. Along the cuesta face in area 6 the competent sandstone capping, as a result of undermining or sapping of the underlying shales, is slumping down the face of the mountain as steep talus slopes. The sandstone is cut by a system of through-going joints producing a coarse columnar structure with individual columns having horizontal dimensions up to 15 or 20 feet. As the individual columns lose their support they topple forward away from the cliff. The columns disintegrate into blocks on falling a' l these blocks constitute the talus. In places the accumulated blocks have moved down the steep mountain face after the manner of rock glaciers but in general they form a talus slope at the foot of the sandstone cliff. The blocks at the lower end of the talus are moved down the mountain face more or less individually as the underlying rocks are eroded.

Block slumping and rock creep.—These two types are associated in areas 1 to 5 but are distinct enough to warrant individual attention. Because both are well displayed in area 3, the discussion which follows refers mainly to that area.

Area 3 consists of two segments, one characterized by block slumping (3a), the other by rock creep (3b). Segment 3a comprises an amphitheater-like feature in which the walls are cliffs of Dakota and Fuson and the floor is Lakota covered with slump blocks of Dakota. The slump blocks are very large, some individual blocks having lengths of more than 300 feet, a width of 50 feet, and a thickness of 30 to 40 feet, the thickness of the Dakota rim rock. The mass of sandstone in such blocks is badly broken but with so little displacement between fragments that the bedding can be easily traced through the entire length. In every instance the bedding now dips steeply toward the rim of the amphitheater. It is obvious that these blocks have moved away from the cliff bottom-first, resulting in a backward rotation.

The amphitheater contains more than a hundred such blocks which are readily recognizable. Locally, the blocks are separated by areas of disorganized blocks of Dakota and near the rim there is evidence of considerable talus slumping.

The lower limit of the amphitheater coincides with the outcrop of the Lakota sandstone which extends under the Dakota debris. This results in a steepened gradient over which the Dakota debris has made a magnificent talus slope leading down to the area of rock creep which comprises the lower portion (3b) of the area.

The lower portion of area 3 comprises a lobate area over one mile long and more than a half mile wide extending from the amphitheater to the floor of Salt Canyon. The slope of the surface is variable; the upper one-quarter mile, comprising the talus slope, has a drop of 200 feet; the central portion has a drop of 200 feet in about 0.6 miles; and the lower part has a drop of 280 feet in 0.4 miles. Thus the gradient changes from 800 feet per mile in the upper part, to 330 feet per mile in the central part and finally to 700 feet per mile in the lower part. A noticeable steepening at the terminus accentuates the lobate character of the area.

The central portion of the lobe, just below the talus slope, is relatively free of sandstone debris. That which is present consists of occasional narrow ridges of Dakota and Lakota blocks elongated transverse to the long axis of the lobe. Between these ridges the surface is underlain by Morrison shales and is marked by scattered blocks of Dakota and Lakota. The size and shape of the ridges and the fact that bedding can be traced through some of them indicates that they are old slump-blocks that have retained their identity in spite of their considerable movement. The local relief between ridges and adjacent swales varies from a few feet to as much as 30 or 40 feet. Nowhere is the sandstone debris more than a surface mantle as outcrops of Morrison shales may be observed in the swales wherever the sod has been removed. The lower portion of the lobe contains abundant sandstone debris. On the steep frontal margin of the lobe the sandstone blocks show no characteristic pattern or arrangement but more or less completely mantle the surface of the underlying Morrison.

The nature of this lobe and its segments are shown in the aerial photo, plate 1, and in the cross section, figure 2.

MECHANISM OF THE WASTING

The nature and occurrence of mass-wasting on Table Mountain makes it possible to reach definite conclusions regarding the details of the processes involved. The discussion

which follows will refer mainly to features described in the preceding section with only incidental reference to features elsewhere on the mountain.

Talus slumping.—The mechanism of the talus slumping in area 6 and elsewhere on the mountain is obvious and needs no elaboration. Essential factors are the jointed character of the sandstone, an escarpment high enough to permit the underlying shale to be removed by sapping, and a slope steep enough to permit the removal of the slump material. The slump material must be removed rapidly enough to prevent its accumulation in amounts sufficient to prevent further sapping.

Block slumping.—Block slumping and the associated areas of rock creep are genetically related and together comprise the most significant type of wasting on the mountain. The essential factors for block slumping are: jointing of the sandstone, an escarpment exposing the entire sandstone section and facing down-dip, an adequate supply of water at the shale-sandstone contact to provide lubrication, and removal of the slumped material.

On Table Mountain block slumping has occurred in areas where large, undissected dip slopes were present after streams on the lower slopes had eroded through the capping sandstones. In area 3 the dip slope extended at least from the present north face of the mountain to the vicinity of Salt Canyon. When the stream in Salt Canyon had eroded through the sandstones the environment for mass-wasting was established.

Rain falling on the dip slope would penetrate the Dakota and move down-dip along the contact with the shale. Where conditions were favorable the water provided sufficient erosion and modification of the shale, and perhaps lubrication, to permit the sandstone to slide, carrying portions of the shale with it, in such a manner that the blocks rotated backward as they moved outward and downward from the scarp. Once the Dakota and Fuson had slumped, the water could penetrate the Lakota and produce additional slumping along the contact with the Morrison shales. Thus the Dakota escarpment would maintain a position up slope from the Lakota escarpment as in the upper portion of area 3 at present.

Plate 1. Aerial photo of Table Mountain showing lobate areas 3 and 4 of figure 1 and the upper part of area 1. Courtesy, U. S. Forest Service.

Block slumping would occur in the Lakota but because of the more friable nature of the rock and the presence of the slumped

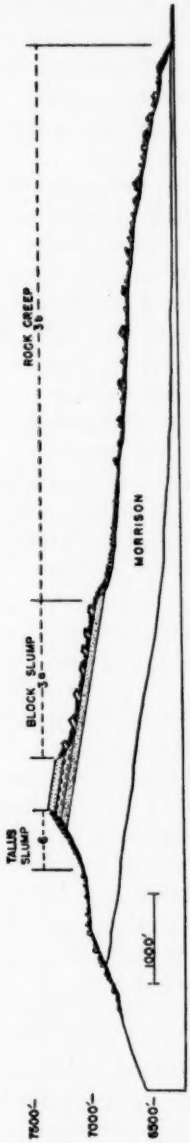

Fig. 2. Cross section of Table Mountain along the line A-B on figure 1.

Dakota debris above, the blocks would show greater disorganization.

Structural sags in the dip slope would serve to concentrate the flow of water in structural lows and thus determine the location of slump areas. There is a broad, gentle sag in the Dakota beds exposed in the head wall of the amphitheater of area 3. We believe this was an important factor in locating the area of slump.

The magnitude of the process is indicated by the fact that more than 600 million tons of Dakota sandstone have slumped in the amphitheater of area 3 alone.

There is no evidence of recent slumping in the amphitheater and trees growing on the slump blocks, some cedars and pines as much as 2 feet in diameter, show no tilting which would result from recent movement. The lack of recent slumping and movements may be due to either or both of the following factors. The limited area of dip slope on the mountain above the area of slumping is too small to provide adequate water for the process of slumping. It is also possible that the slumping was active under a somewhat more humid climate of the past. Either factor would be sufficient to explain the lack of recent slumping but we are inclined to favor climate as the critical factor.

Rock creep.—It is obvious that the slump debris is being transferred down the slope of the lobe and, for want of a better name, we are calling this transfer rock creep. The essential cause of the down-slope movement is the erosion of the Morrison shales. These are soft, poorly indurated, clay shales which become extremely slippery when wet and offer essentially no resistance to erosion. Drainage channels extend upward on the lobe to the foot of the talus slope just below the amphitheater. Evidence of rapid erosion is present almost everywhere and as the shales are removed the slump debris is caused to move down the slope. The decreased friction when the shales are wet would further accelerate the movement.

The central portion of the lobe is relatively free of slump debris because the material has continued to move down the slope whereas the supply of additional material from above has effectively ceased. Under conditions of more active erosion of the shales in the past the Lakota sandstone was being continually undermined by the removal of the Morrison shales.

This permitted the Lakota to slump downward carrying with it the already slumped Dakota material. This slumping was, at least in part, of the block type as evidenced by the concentration of debris in transverse ridges in the area of rock creep. In these ridges of debris Dakota and Lakota fragment are intermingled. Because of the poor cementation of the Lakota the fragments of it disintegrate more rapidly and this accounts for the great predominance of Dakota debris on the lower slopes of the lobe.

Debris fragments which remain intact during the transfer of the lower end of the lobe disappear beneath the finer alluvium in Salt Canyon. In general they should form a mantle on the bedrock surface of the valley to be exposed only in deep gullies cut after severe storms. In such an environment they would continue to disintegrate until entirely destroyed and removed as sand.

The processes which operate in the area of rock creep would also be influenced by the same factors discussed with respect to block slumping. The mechanism requires a supply of water, especially at the upper end of the creep area. This could be supplied by a larger catchment area on the mountain or by greater rainfall. The lack of continued slumping of the Lakota has stopped the supply of debris to the area of creep which explains the scarcity of debris in the upper portion of that area. For area 3 the whole process of mass-wasting seems to have been effectively halted except for slow creep at the lower end of the lobe.

Areas 4 and 5 are lobate areas very similar to area 3 except that the eastward extension of Table Mountain across the heads of these areas has been entirely destroyed. If the major factor in the slumping is an adequate catchment area then some explanation must be found for the eventual destruction of the mountain top at the head of such a lobe after the catchment area reaches an effective minimum. This could be accomplished by talus slumping on the north face of the mountain but this fails to account for the continued effective action in the area of creep on each lobe. We believe that all aspects of the mass-wasting on the mountain can best be accounted for on a basis of a more humid climate in the past. All of our observations indicate that we see here essentially the final stages of a process stopped by a decrease in rainfall.

Under adequate rainfall the decrease in catchment area would slow the process of mass-wasting but areas 4 and 5 are evidence that the catchment area is not the critical factor.

MASS-WASTING AND EROSION

It is obvious that mass-wasting has been a major factor in the erosion of Table Mountain. Here, under a combination of favorable factors, a mountain of large size and relief has been largely destroyed. This type of erosion has received too little attention as evidenced by the general idea that erosion is a slow process. Although mass-wasting on Table Mountain is developed to a spectacular degree it can be and is no less effective elsewhere. The whole region around Table Mountain furnishes abundant evidence of the collapse, by mass-wasting, of the cuesta-like ridges of Dakota and the underlying Lakota sandstones. The net effect of these processes on the rate of erosion must be carefully assessed.

Observations such as recorded here have major implications for our thinking about the erosion of a mountain range such as the Colorado Front Range. As the great structural arch of the range began to rise and erosion was initiated there must have been many situations favorable for mass-wasting of the sedimentary rocks. Such areas were not only numerous but large, thus making it entirely possible that mass-wasting was a major factor in the removal of the sedimentary rocks. If we are permitted an analogy between Table Mountain and the early range then it would seem that mass-wasting may have been a factor in the establishment of drainage lines in the area. It may be that many, so-called, consequent streams actually obtained their courses as a result of mass-wasting.

Of special interest are the implications of these processes in any analysis and interpretation of the high-level erosion surfaces (peneplains) of such a mountain region. Mass-wasting would not only influence the development of drainage lines and the development of valleys but would greatly influence the average rate of erosion and thus the time required to produce an extensive erosion surface. The effect of the abundant coarse debris on the activity of streams might even favor the development of extensive erosion surfaces with

unusually steep gradients which would resemble a conventional peneplain only in being flat.

An examination of Table Mountain in the field is recommended for anyone who is interested in this process. The area is conveniently located and one can examine the pertinent features in a period of hours. A visit to the area is well worth the time and effort required.

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TOPOGRAPHIC DISCONTINUITIES OF THE DES MOINES LOBE*

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ABSTRACT. Although the Cary drift of the Des Moines lobe is differentiated from the older adjacent drifts by four lines of evidence, only one, erosional topographic discontinuity, has universal areal application from the head of the Coteau des Prairies in eastern South Dakota around the lobe to Dakota County, Minnesota. This criterion does not apply to differentiation of the Cary of the Des Moines lobe from the younger adjacent Mankato drift of the lobe nor from the younger adjacent Cary red drift in east-central Minnesota.

Angular topographic discontinuity sets off the Mankato drift from the Cary drift of the Des Moines lobe at the margin of the Altamount moraine.

Detailed mapping of the Bemis and Altamount morainic systems around the Des Moines lobe and northward into Dakota County, Minnesota shows that the Bemis system (Cary) is overlapped by the Cary red drift of east-central Minnesota and that the Cary red drift is, in turn, overlapped by the Altamount morainic system (Mankato) of the Des Moines lobe.

INTRODUCTION

THE morainic complexes of the Des Moines lobe in Iowa and Minnesota long have been recognized by geologists familiar with this region. White (1870, pp. 98-99) recognized the multiplicity of the moraines in Iowa. Chamberlin (1883, pp. 388-393; plates 28, 35) described two major systems (Altamont and Gary) on the eastern flank of the Coteau des Prairies in eastern South Dakota and southwestern Minnesota. These two systems were delineated also in Iowa. Upham (1884, pp. 406, 479, 505-521, 602-606) described and located these features in Minnesota and Iowa. Leverett (1922) demonstrated that the Altamont moraine of Chamberlin was not the marginal moraine of the Des Moines lobe but that the Bemis, several miles to the west of the Altamont in eastern South Dakota, constituted the margin of the lobe. The term Altamont was maintained for the second major system. Smith (1925, opp. p. 102) delineated a third and fourth system in Iowa that he designated the Humboldt and Algona. However, he used Chamberlin's terminology and named the first and second systems the Altamont and Gary. Leverett (1932, pp. 57-61, 67-72) reported on earlier field studies with Lees in Iowa and renamed the first and second systems the Bemis

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and the Altamont. Kay and Graham (1943, p. 239) recognized four major systems in Iowa from the margin of the lobe inward, the Bemis, Altamont, Humboldt, and Alogna. This sequence is generally accepted in Iowa.

With one exception, the Des Moines lobe has always been considered one drift, the Wisconsin of earlier classifications, the Late Wisconsin of Leverett, and since 1933 (Leighton, p. 168), the Mankato. The exception is an interesting map published by Atwood (1940, pp. 198-199) that is accredited to William C. Alden.[1] A morainic margin of an ice sheet that roughly coincides with the Altamont moraine is shown on the sketch map to subdivide the Des Moines lobe into two drifts.

During 1948 and 1949 field study of the Des Moines lobe was conducted by the present writer in northwestern Iowa and southwestern Minnesota for the Iowa Geological Survey. A preliminary report of the results of this study has been presented (Ruhe, 1950b). During the summer and fall of 1950 the work was continued in north-central Iowa and south-central Minnesota under the auspices of a National Research Fellowship in the Natural Sciences. Field expenses were subsidized by the Iowa Geological Survey for the study in Iowa and by the Industrial Science Research Institute, Iowa State College for the study in Minnesota. Topographic criteria are discussed in this paper that confirm: (1) differentiation of the Des Moines lobe from adjacent drifts, and (2) the twofold division of the Des Moines lobe at the margin of the Altamont moraine.

DIFFERENTIATION OF THE CARY DRIFT

The Cary drift of the Des Moines lobe is defined as that part of the lobe from the margin of the lobe to the margin of the Altamont moraine (M-1, fig. 1), and includes all of the Bemis morainic system (C-1, C-2). In general, four criteria differentiate the Cary drift from the older adjacent drifts. These criteria are: (1) prominent marginal end moraine, (2) a thin discontinuous, poorly sorted loess mantle on the Cary drift in contrast to a continuous, well-sorted

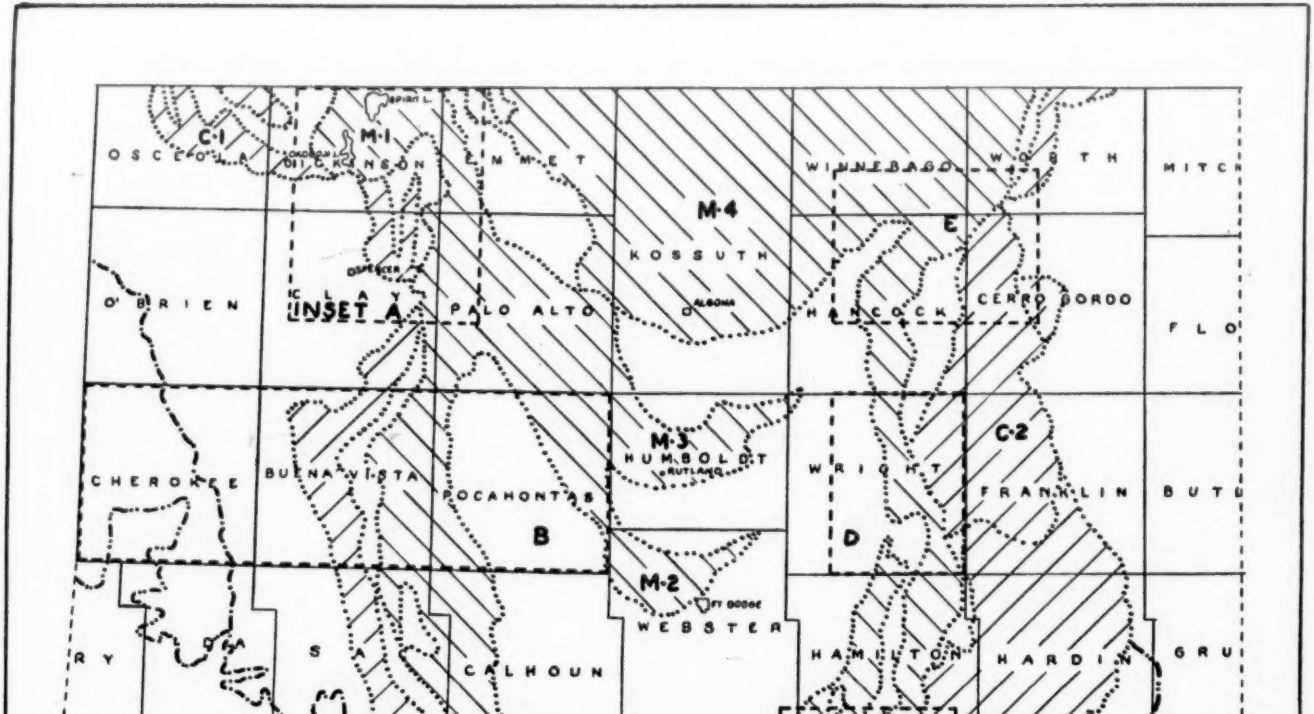
[1] The map was brought to the attention of R. F. Flint who traced its history. The map is a copy of an unpublished map that was compiled by W. C. Alden and F. E. Matthes and is based in part on the field work of Alden, Leverett, and other geologists.

loess mantle on adjacent older drifts, (3) exposed sections showing calcareous Cary till on well-sorted Iowan-Tazewell loess or older deposits and (4) erosional topographic discontinuity at the margin of the Cary drift.

A prominent marginal end moraine exists on the lobe in eastern South Dakota and trends southeastward in Minnesota through Lincoln, Pipestone, Murray, and Nobles Counties. This moraine (C-1, fig. 1; Bemis moraine) enters Iowa in north-central Osceola County and can be traced around the lobe to the Minnesota line in Worth County, Iowa. From Worth County the Bemis moraine trends northeasterly through Freeborn County into Mower County, Minnesota thence northward through Dodge and Rice Counties to the St. Croix moraine of the Patrician red drift in Dakota County, Minnesota. The marginal end moraine is not prominent, however, at many places along this traverse. For example, in Dickinson and Clay Counties, Iowa the moraine rises only 5 to 10 feet above the adjacent loess-mantled Tazewell ground moraine. Lineation of ridges within the morainal system is determinable on aerial photographs but is difficult to check in the field because of the low relief. Placement of the margin of the moraine requires the use of the other criteria of differentiation.

From Carroll, Iowa southeastward through Des Moines and northeastward to the Marshall County line a prominent marginal end moraine is not evident. In this area the Cary drift is in juxtaposition to the loess-mantled Kansan drift. In Cerro Gordo County, Iowa the small northeastward-trending Cary lobe does not stand appreciably above the adjacent Iowan drift .

The second criterion, thin, discontinuous, poorly sorted loess mantle on the Cary drift in contrast to a continuous well-sorted loess mantle on adjacent older drifts, has wide regional application. It delineates the Cary drift from the adjacent Tazewell in southwestern Minnesota and northwestern Iowa. The Cary is also differentiated on this basis from the Kansan drift, with its thick, well-sorted loess mantle, from Carroll County to east-central Hardin County, Iowa. Northward to southern Cerro Gordo County, Iowa the loess-mantled Iowan drift is adjacent to the Cary. However, from Cerro Gordo County northward through Worth County, Iowa, and


C-1
M-1
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E
INSET A
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B
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M-2
O'BRIEN
CLAY
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KOSSUTH
WINNEBAGO
WORTH
MITCH
HANCOCK
CERRO GORDO
FLO
CHEROKEE
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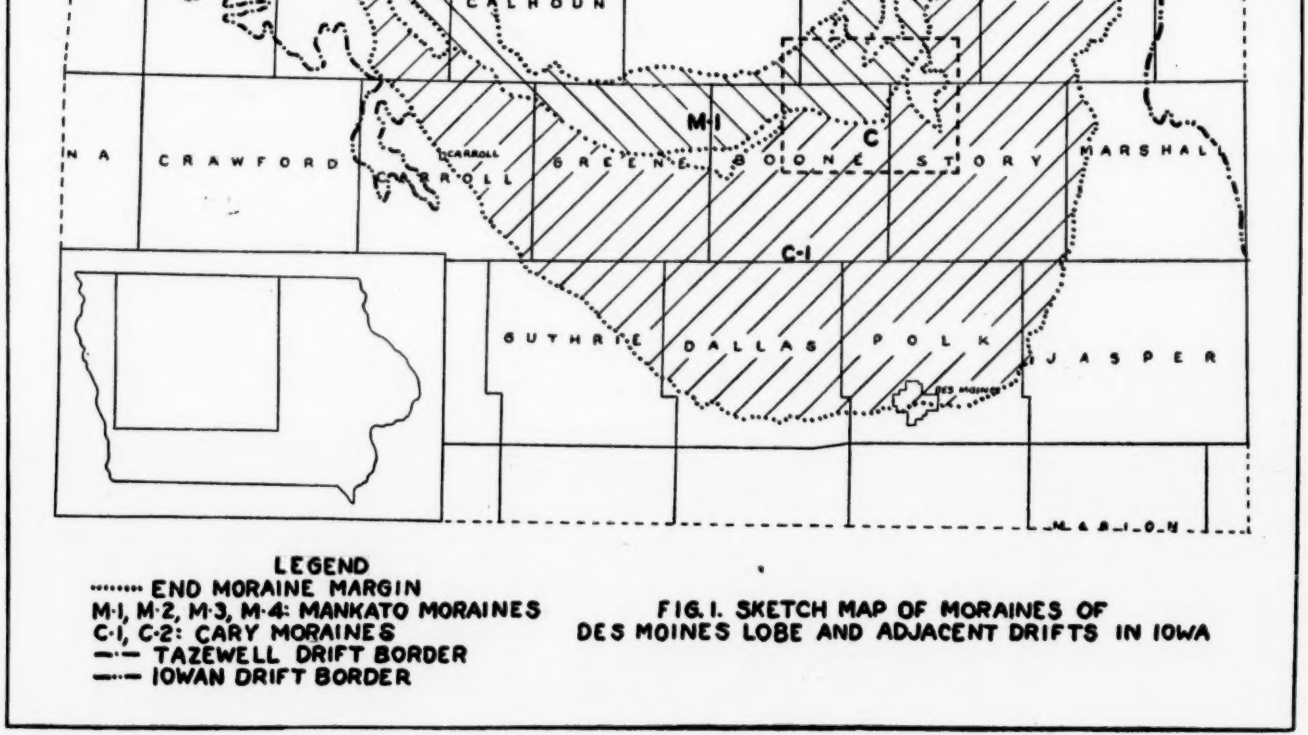

FIG. 1. SKETCH MAP OF MORAINES OF DES MOINES LOBE AND ADJACENT DRIFTS IN IOWA

Fig. 1

Freeborn, Mower, Dodge, Rice, and Dakota Counties, Minnesota the Iowan drift is discontinuously covered by a thin poorly sorted loess that is similar to the loess on the Cary.

Although stratigraphic sections showing calcareous Cary till over well-sorted Iowan-Tazewell loess constitute the proof of an advance of Cary ice, they are unimportant in regard to the mapping of the geographic distribution of the younger drift. Sections are so few and so widely separated geographically that mapping of limits of a drift on this basis would be impossible.

EROSIONAL TOPOGRAPHIC DISCONTINUITY

Only the fourth criterion, erosional topographic discontinuity, holds true continuously around the Des Moines lobe from South Dakota to Dakota County, Minnesota. Erosional topographic discontinuity is referred to in this paper as the difference in degree of integration of drainage on the surfaces of adjacent drifts. The degree of integration of drainage on a drift surface is a reflection of the erosional modification of that surface by subaerial processes after it had been uncovered, but not overridden by a subsequent advance of glacier ice. The poorly drained Cary surface is in marked contrast to the older well-drained surfaces that are adjacent to it. The well-drained Tazewell surface contrasts sharply with the poorly drained Cary surface in southwestern Minnesota and northwestern Iowa. Undrained depressions are abundant on the Cary surface, but are either rare or entirely lacking locally on the Tazewell surface.

A detailed drainage map that includes parts of the Kansan, Iowan, Tazewell, Cary, and Mankato surfaces is shown in figure 2. The map encompasses the area of inset B of figure 1. Well-integrated patterns are shown on the Iowan and Tazewell surfaces and poorly integrated patterns on the Cary and Mankato surfaces. Three significant facts are brought out by this map: (1) drainage patterns of the Iowan and Tazewell surfaces are similar; (2) drainage patterns of the Cary and Mankato surfaces are similar; and (3) a marked contrast is noted between the well-integrated Tazewell system and the poorly integrated Cary system. Drainage relationships of the Wisconsin drift surfaces support previous conclusions (Ruhe,

1950a, pp. 441-442; Flint, 1950a) that the Iowan and Tazewell are closely related in time and that the Cary and Mankato are a second closely related pair. The contrast of the Cary and Tazewell surfaces implies that the Tazewell-Cary interval was the major time break in the Wisconsin sequence and was long enough to allow establishment of drainage on the Tazewell surface prior to Cary glaciation.

The contrast of the Cary drainage with an older surface illustrated in figure 2 is characteristic around the Des Moines lobe to the area of the Patrician red drift in east-central Minnesota (Leverett, 1932, pl. 2). An erosional topographic discontinuity separates the poorly drained Cary surface from the well-drained loess-mantled Kansan surface around the southern part of the lobe from Carroll to Hardin County, Iowa. Northward to Dakota County, Minnesota the poorly drained Cary surface is differentiated from the well-drained Iowan surface on the same basis.

In Dakota County, Minnesota the Cary surface of the Des Moines lobe and the Cary red drift surface are very similar topographically. Both are poorly drained. Surface drainage is not integrated; undrained depressions are abundant. Both surfaces are end morainic. The Cary red drift is typically a knob and kettle complex. (See St. Paul quadrangle topographic sheet.) The Cary of the Des Moines lobe is characterized by linear ridges that trend north-south.

At several places in southwestern Dakota County the linear end-morainic ridges of the Des Moines lobe are locally capped by Cary red drift. Thus the marginal part of the Des Moines lobe (the drift of the Bemis morainic system) is slightly older than the Cary red drift of east-central Minnesota and passes under the red drift in Dakota County, Minnesota.

In summary, erosional topographic discontinuities separate the Cary drift (Bemis morainic system) of the Des Moines lobe from adjacent older drifts from the head of the Coteau des Prairies in eastern South Dakota around the lobe to Des Moines, Iowa and northward to Dakota County, Minnesota. This criterion is not applicable to a differentiation of the Cary drift of the Des Moines lobe from younger drifts, the Cary red drift of Minnesota or the Mankato drift of the Des Moines lobe.

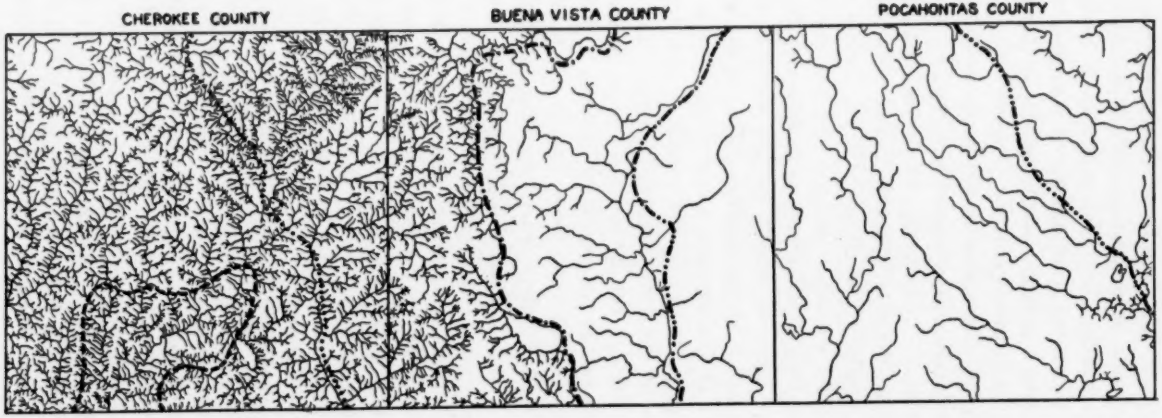

FIG. 2. TOPOGRAPHIC DISCONTINUITY AT MARGIN OF CARY DRIFT IN NORTHWESTERN IOWA

Fig. 2

ANGULAR TOPOGRAPHIC DISCONTINUITY

Detailed mapping of the trends of the crests of end-morainal ridges within a morainic system[2] shows that one system may overlap another and demonstrates a readvance of glacier ice. The discordance of trends of morainal ridges of two adjacent morainal systems is referred to in this paper as angular topographic discontinuity.

Angular topographic discontinuity exists in many places between the Altamont morainic system (M-1, fig. 1) and the Bemis morainic system (C-1, C-2, fig. 1) of the Des Moines lobe. In southwestern Roberts County, South Dakota the Altamont moraine overlaps the Bemis (Flint, 1950b). To the southeastward these moraines diverge and are separated by a ground-morainal belt 25 miles wide in Nobles and Jackson Counties, Minnesota. Leverett (1932, p. 69) recognized this divergence and explained it by the buttress effect of the prominent quartzite range in Cottonwood County, Minnesota.

In Dickinson County, Iowa, to the south of the area of maximum divergence of the two morainic systems, the Altamont moraine also overlaps the Bemis. (See inset A, fig. 1.) Discordance of trends of the two morainal systems is distinct. Trends of the Bemis moraine (C-1 in A, fig. 3) are east-west to the south end of Lake Okoboji where they veer directly to the south into Clay County. Trends of the Altamont moraine (M-1) form a lobe around Okoboji and Spirit Lakes and truncate the Bemis moraine at the south end of Lake Okoboji. Overlap indicated by angular topographic discontinuity of the Altamont and Bemis systems is shown also in inset A, figure 3, approximately 9 miles east of Spencer, Iowa. In this area the Mankato drift completely overlaps the Cary and is in contact with the Tazewell.

From this area southward to northern Boone County, Iowa the Altamont moraine in general parallels the Bemis moraine. Angular discontinuity of trends of these moraines exists in northern Boone County (Inset C, figs. 1 and 3). The Bemis (C-1) is overlapped and truncated by the Altamont (M-1).

Angular topographic discontinuity exists in eastern Wright County, Iowa (Inset D, figs. 1 and 3). Trends of the Altamont system (M-1) truncate those of the Bemis system (C-2)

[2] R. F. Flint, while a member of the U. S. Geological Survey, used this mapping technique in his study of the Pleistocene of South Dakota.

at nearly a right angle. Farther to the north in Cerro Gordo, Hancock, Winnebago, and Worth Counties, Iowa, angular discontinuity of these two systems also exists (Inset E, figs. 1 and 3).

The angular discordance of trends of morainic ridges of the Altamont and Bemis systems shows the overlap of the older by the younger complex. Overriding of the older moraines by a younger ice mass is indicated. This evidence appears to modify an earlier interpretation (Kay and Graham, 1943, p. 242) which is here quoted:

> "The question as to the approximate contemporaneity of these various moraines of the Des Moines lobe has never become an issue due to the

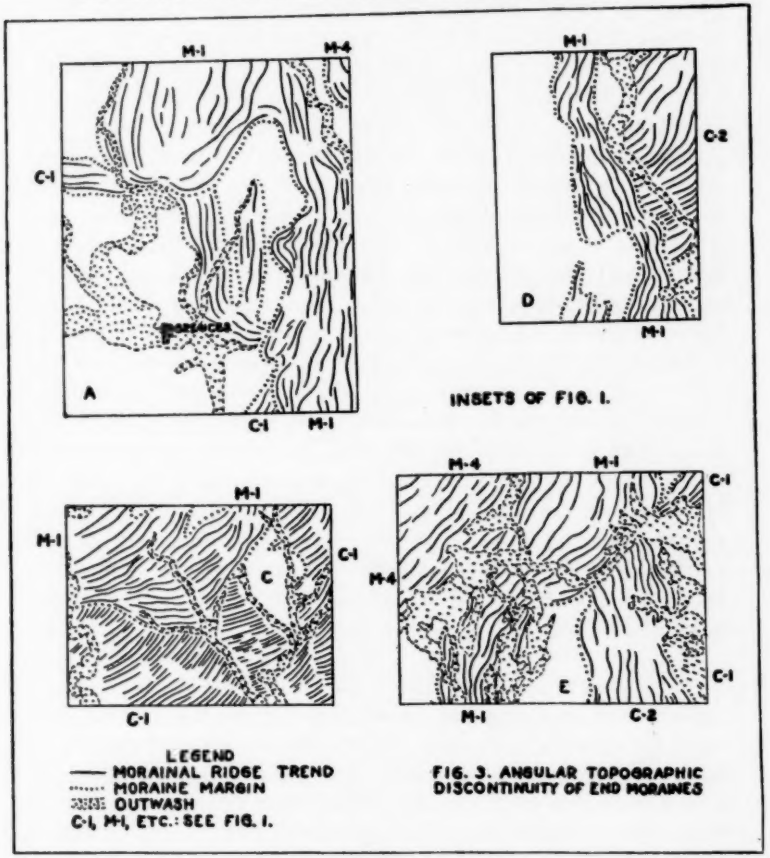

Fig. 3

> obvious array of evidence pointing to normal stages in the advance and recession of a single glacial mass. In morainal characters as to material, topography, position, and reduction by erosion; in appearance, and relation of outwash materials to drainage features; in degree of surface till modification by weathering of the various areas of the lobe, and in physiographic age of the various areas of the lobe, the similarities are so striking and the unity of the effects of glaciation so apparent, that glacial geologists have in general accepted the single glacier view of the Mankato [Des Moines lobe] drift in Iowa and its morainic complex."

Further evidence that demands abandonment of the single glacier concept is found in Dakota County, Minnesota. The Bemis moraine (Cary) was noted in an earlier part of this paper to trend northeasterly from Worth County, Iowa through Freeborn and Mower Counties, Minnesota to Austin where the trend changes to northerly through Mower, Dodge, and Rice Counties into Dakota County where this drift was found to pass under the Patrician red drift (Cary). The Altamont moraine (Mankato) was traced by the writer northward from Worth County, Iowa through Freeborn, Steele, and Rice Counties, Minnesota into Dakota County where this drift was found to overlap the Patrician red drift (Cary). Leverett (1932, pp. 56-78) and Sardeson (1932, pp. 78-90) reported many localities where the Wisconsin gray drift [Mankato] of the Grantsburg sublobe is superimposed on Wisconsin red drift [Cary] in this area. They believed, however, that the outer system of moraines in the sublobe, the Rush Lake, was correlative with the Bemis morainic belt. The present writer believes that the drift of the Grantsburg sublobe is a correlative of the drift of the Altamont morainic system. Tracing of moraines on the Des Moines lobe shows that the Altamont rather than the Bemis moraine constitutes the marginal moraine of the Grantsburg sublobe. The drift of the Bemis moraine passes under the Wisconsin red drift in the southwestern part of Dakota County.

TOPOGRAPHIC CONTINUITY

Erosional and angular topographic discontinuities do not exist between the morainic systems of the Mankato drift in Iowa. The Humboldt moraine (M-2, M-3, fig. 1) develops as a bifurcation of the broad Altamont-Humboldt complex in Palo Alto County, Iowa. To the eastward the southern phase of the Humboldt, the Fort Dodge moraine, grades into the

Mankato ground moraine in northeastern Webster County. The northern phase of the Humboldt system, the Rutland moraine, grades into the Mankato ground moraine in northeastern Humboldt County and northwestern Wright County. The Humboldt system is believed to represent only a recessional phase of Mankato history.

The Algona moraine (M-4, fig.1) of northern Iowa (Gary of Leverett in Minnesota) develops as a bifurcation of the Altamont-Gary system in Lincoln County, Minnesota and merges again with the Altamont system with apparent continuity in eastern Winnebago County, Iowa. However, the massiveness and vigorous topographic expression of this morainic belt in northern Iowa suggest that the system may represent more than just a phase in the general recession of the Mankato ice. Lack of evidence at the present time relegates the Alonga system to a status similar to the Humboldt system in the interpretation of the Mankato glacial history.

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ORBICULAR DIORITE FROM SOUTHERN ALASKA*

RICHARD G. RAY

ABSTRACT. Quartz diorite in the Willow Creek mining district of southern Alaska locally contains orbicular structures as much as 10 inches across. The orbicules are found mostly in loose blocks of quartz diorite adjacent to a pluglike mass of andesine pegmatite within the Talkeetna batholith. It is inferred that the pegmatite-quartz diorite contact was the locus of orbicule formation.

Minerals similar to those of the host quartz diorite occur in conspicuous concentric bands or shells in the orbicules. Radial growth of plagioclase is prominent in certain of the shells.

Deformation effects, such as contortion of orbicule shells and actual rupture of some shells, show that the matrix could not have been solid when the orbicules formed. The orbicules began to develop sometime after the host quartz diorite had started to crystallize but were completely formed prior to the final consolidation of the host rock. Genesis of the orbicular structures is discussed only so far as the environment of formation is concerned. No attempt is made to explain the concentric and radial nature of growth.

INTRODUCTION

ORBICULAR rocks were first described more than 100 years ago, but the number of occurrences known today is still very small. Although most orbicules have been found only in glacial boulders or other loose rocks, they have received wide notice because of their geologic rarity and their beauty, and particularly because of their possible petrogenic significance. A number of hypotheses have been offered to explain their origin, but it is doubtful that any one hypothesis should be rigidly applied to all orbicule occurrences.

From a study of the orbicular granite at Kortfors, Finland, Bäckström is said to have developed the "liquation hypothesis" (see Eskola, 1938, pp. 466-467 and Sederholm, 1928, p. 48), proposed earlier by Durocher. Accordingly, orbicules formed as secretions from immiscible or partially miscible drops of one magma within another. Laboratory studies have since shown that most silicate melts are miscible, and the liquation hypothesis is now generally discredited.

As early as 1896 Frosterus questioned the secretionary nature of orbicule formation (1896, p. 34). From a study

* Publication authorized by the Director, U. S. Geological Survey.

of the well-known Kangasniemi orbicular granite he concluded that the orbicules formed around foreign nuclei within magma zones of peculiar composition.

Some investigators have considered orbicules as assimilated inclusions; but it is well known that inclusions are common in plutonic rocks, whereas orbicules are very rare.

Both Watson (1904) and Lawson (1904) have described orbicules at Davie County, North Carolina, and at Dehesa, California, respectively, as products of magmatic differentiation. Sederholm (1928) also considered orbicular structures as products of magmatic differentiation, but the chemical composition of the orbicules led him to postulate a special magma type.

More recently the formation of orbicular structures in certain Finnish terranes has been attributed to ultrametamorphic processes by Eskola (1938) and Simonen (1940, 1950). The close chemical-mineralogical relationship between orbicules and a secondary matrix formed by granitization led Eskola to postulate their formation by metamorphic differentiation according to the concretion principle. Such structures are thought to have developed in a nearly solid rock. The migmatitic nature of the terrane in which most of the Finnish orbicules occur lends support to the hypothesis that some orbicules formed by metamorphic differentiation and replacement.

De Jong (1943) has concluded that the orbicules found at Eext, the Netherlands, formed by replacement processes shortly after the solidification of an original magma.

Although there has been considerable support recently for a secondary origin of orbicules, the possible primary origin of certain of these structures, especially those occurring in nonmigmatitic rocks, cannot be disregarded. The orbicules from southern Alaska are in a nonmigmatitic environment. They were discovered during the summer of 1949 while the writer was engaged in mapping a marginal segment of the Talkeetna batholith. Although orbicular rocks have been found in many parts of the world, none have been described from Alaskan localities, as far as the writer is aware, with the exception of some crude "orbules" discovered by Knopf at a granite-limestone contact on Seward Peninsula (1908, pp. 45-49). These are poorly developed structures and are dissimilar to the ones discussed herein.

LOCATION AND GEOLOGIC SETTING

The orbicular rocks described below were discovered in an area of rugged relief in the Willow Creek mining district of southern Alaska about 20 miles north of the town of Palmer and 70 miles north of the city of Anchorage. Quartz diorite underlying the district is rather uniform mineralogically over an area of at least 40 square miles. It is characterized by a moderately well developed primary flow structure which has a constant trend to the northeast and moderate dip to the northwest. Hundreds of small mafic inclusions are present in the quartz diorite. These inclusions are similar mineralogically to the host quartz diorite. They differ in having a higher percentage of mafic minerals. Such inclusions even at altitudes more than 1000 feet below the orbicule locality—deeper in the batholith—rarely show signs of corrosion or reaction with the quartz diorite. The only other rocks in the vicinity of the orbicule locality are younger conglomerate, sandstone, and arkose, which lie unconformably on the batholith front. Most of the orbicular structures are in loose blocks of quartz diorite on a spur nearly a mile northeast of the Snowbird mine camp. Blocks of quartz diorite containing the orbicules are localized in an area about 20 feet square. This locality is 3 miles north of the southern border of the Talkeetna quartz diorite batholith. Judging from the numerous mafic inclusions in the surrounding quartz diorite, it is probably only a short distance below the original roof of the batholith, now eroded away. There can be no doubt that the orbicules are within the batholith, which is generally considered to be late Mesozoic in age.

The largest block of quartz diorite containing orbicules measured approximately 10 feet by 5 feet on the exposed face. No blocks were found in place, but because of their large size and angular form, and their position only 40 to 50 feet below the crest of a spur, it is unlikely that they have been transported any distance, but have probably tumbled only a few feet out of place. The quartz diorite of the area has been intruded by a small mass of andesine pegmatite not more than 50 feet across, apparently pluglike in plan as it could not be traced in any direction. Quartz diorite boulders containing orbicules lie adjacent to the pegmatite.

MEGASCOPIC FEATURES

The orbicules are triaxial ellipsoids with short and long axes ranging from 1 by 1½ inches to about 6 by 10 inches. Most ellipsoids are separated by a matrix of normal quartz diorite in which they show a tendency towards orientation of the long axes. A conspicuous concentric banding of minerals similar to those in the quartz diorite matrix characterizes the orbicules (plate 1).

Orbicular structures of any one loose slab of rock tend to be of the same size—either all large or all small. The small orbicules, mostly about 2 to 2½ inches in longest diameter, were found only in one block of pegmatitic material. Within the small orbicules four distinct mineral zones are recognizable. From the center to the rim these are:

1. A nucleus consisting of a crystal of andesine 0.5 to 0.6 inch across.
2. A mixed zone about 0.4 to 0.5 inch across composed of scattered small crystals of biotite in the white andesine. Only a crudely concentric pattern is shown.
3. A concentric band rather consistently about 0.4 inch wide composed almost entirely of white andesine.
4. An outer skin of mafic minerals, principally biotite tangentially arranged.

The central parts of the large orbicules correspond almost identically in mineral make-up and width of concentric bands to the small orbicules described above. From the center towards the rim the following zones, in addition to the first three zones described above, are present in the large orbicules:

4. A zone of plagioclase containing scattered mafic minerals. This mixed zone is about 0.6 to 0.8 inch wide along the long axis of the orbicule. It is not uniform in mineral distribution but shows a concentration of mafic minerals, largely biotite, at the outer and inner borders of the zone.
5. A conspicuous concentric band generally between 0.10 and 0.15 inch in width composed almost entirely of white andesine.
6. A dark colored zone 1 to 2 inches wide containing several thin but conspicuous black shells of biotite.
7. An outermost dark-gray zone as much as 2 inches wide composed of a mixture of fine-grained andesine crystals and biotite.

MICROSCOPIC FEATURES

Nucleus.—Weakly zoned plagioclase having an anorthite content ranging from 38 to 41 per cent forms the orbicule nuclei. Multiple twinning parallel to the side pinacoid and parallel or nearly so to the basal pinacoid is strongly developed. Generally only a single plagioclase crystal 0.5 to 0.6 inch across forms an orbicule nucleus, and this is similar in size, anorthite content, and manner of twinning to plagioclase crystals in adjacent pegmatite.

Shells.—The feldspar of the orbicule shells is almost entirely plagioclase like that of the surrounding quartz diorite, but it exhibits an overall weak zoning, whereas plagioclase grains in the quartz diorite matrix generally show a weakly zoned core separated by a strong compositional break from a weakly zoned outer rim. The plagioclase from four zones within the orbicules—namely the nucleus, each of the white feldspar shells, and the outermost fine-grained shell—was examined by the immersion method. Averages of many readings to determine the lowest α and highest γ indices in the various zones gave the following results:

	Per cent anorthite
Nucleus	38 - 41
Inner white feldspar shell	37 - 40
Outer white feldspar shell	35 - 37
Outermost fine-grained shell	35 - 37

This range is smaller than that of the plagioclase of the quartz diorite, which is between 35 and 47 per cent.

Radial growth of plagioclase crystals, especially in the white shells composed almost entirely of plagioclase, is well shown in sawed and polished slabs. In thin sections at least three distinctive habits of the feldspar are observed. One growth habit exhibits elongate crystals slightly radially directed, but subparallelism of these grains is a dominant characteristic. A second growth habit shows elongate plagioclase crystals which are strongly divergent, although the radial nature of growth is preserved. A similar growth habit in quartz has been described as "flamboyant" or "feathered" (Adams, 1920, pp. 628-629, 657). Besides these two distinctive growth habits of elongate crystals, the plagioclase in places, particularly in the fine-grained outer orbicule shell, occurs merely as small equidimensional grains randomly ori-

ented—giving that part of the orbicule a granular texture. In the white bands composed almost wholly of plagioclase one or the other of the radial habits usually predominates; whereas in darker-colored, more mafic zones equigranular feldspar crystals generally are most common. However, in some places all three habits are present.

Multiple twinning of the plagioclase is common parallel to the side pinacoid and parallel or nearly so to the basal pinacoid. Twin types have been determined by universal stage methods and include: albite and albite-Ala B twins parallel to the side pinacoid; and acline or pericline twins parallel to the basal plane.[1] These twin types all exhibit narrow twin lamellae similar to what is generally referred to as "albite twinning," and examination by universal stage methods is necessary in distinguishing them. In thin sections cut along the radial direction of growth, traces of both the side pinacoidal and basal twin planes may be parallel or somewhat divergent. In contrast, thin sections cut at right angles to the radial direction show a preponderance of traces of side pinacoidal and basal twin planes intersecting at nearly right angles, and thus the *a* axis of the plagioclase is the axis of elongation.

Although biotite and hornblende occur in approximately equal amounts in the quartz diorite country rock, biotite greatly predóminates over hornblende in the orbicules. The biotite generally has formed with the cleavage direction parallel or subparallel to the concentric banding, although in a few places, especially in the outer shells, the coarser biotite plates may be oriented with the cleavage direction at right angles to the concentric banding. Hornblende occurs only sparingly in the orbicules. Where it was observed in one mafic shell the *b* and *c* axes were oriented essentially parallel to the concentric banding.

Epidote is typically confined to the mafic shells where biotite predominates. It is most often seen as isolated patches or as stringers along the cleavage of biotite that is altering to chlorite. Some chlorite grains in turn show stringers of sphene along cleavage planes. In the more mafic zones sphene and

[1] In plagioclase of the compositions here considered (An_{35-47}) the "rhombic section," which is the composition plane of pericline twinning, is so nearly parallel to the basal pinacoid, which is the acline composition plane, that these two twin types are not readily distinguished.

calcite occur sparingly as small isolated masses. Apatite is present but is less abundant in the orbicules than in the surrounding quartz diorite matrix. Rarely microcline is present in the shell adjacent to the orbicule nucleus.

QUARTZ DIORITE MATRIX AND NATURE OF CONTACT WITH ORBICULES

Although a radial structure characterizes the orbicules in large part, the outermost concentric shell is fine grained and has a granular texture. This fine-grained shell gives way abruptly to the medium grain-size of the quartz diorite matrix which has a similar granular texture. At this contact, however, quartz is conspicuous in the matrix material, hornblende occurs in greater abundance than it does within the orbicule, and the plagioclase exhibits a strong compositional break between a weakly zoned core and weakly zoned rim.

The quartz diorite matrix is very similar to the quartz diorite found throughout the surrounding area. A comparison of micrometric analyses of quartz diorite between orbicules with quartz diorite from other localities in the district shows only small differences in the amounts of plagioclase, quartz, hornblende, and biotite. The general range of constituents by volume is:

	Per cent
Plagioclase	60 - 70
Quartz	15 - 20
Biotite	8 - 10
Hornblende	8 - 10
Accessory minerals	1 - 2

Accessory minerals include apatite, sphene, epidote, and in some specimens, microcline. The plagioclase ranges from 35 to 47 per cent anorthite in composition; thus the matrix material is quartz diorite according to the Johannsen classification. It is apparent from a microscopic examination of the orbicules that they are generally similar mineralogically to the quartz diorite matrix, although the proportions of the various minerals differ. The orbicules are notably different from the quartz diorite matrix in containing no primary quartz. Because of the mineral segregation into shells, micrometric analyses of the orbicules would be difficult to make, but it is clear that the orbicules should be classified as diorite.

DEFORMATION OF THE ORBICULES

Deformation structures in the orbicules give the strongest evidence of the environment in which the orbicules formed. The concentric bands have been disturbed in a variety of ways. They may be thickened along the long axis of an ellipsoid; they may be ruptured; or they may be merely contorted. Plate 1, and plate 2, figure 1 show clearly that deformation took place at various times during the formation of the orbicular structures. Figure 1 of plate 2 shows ruptured mafic shells halfway between the rim and center of an orbicule. The ruptured part has been torn completely away and an outer uniform shell has been deposited unconformably on the ruptured surface. At least one stage of deformation thus took place prior to the formation of the wide outer shell. Mutual interference of the orbicules shown in plate 1 has deformed especially the outer shells. The figures of plate 1 show orbicules from which the outer shell, as well as inner shells, have been torn loose, and in figure 1 parts of a ruptured outer shell lie isolated in the quartz diorite matrix. This figure also shows a mafic portion of an orbicule shell which was isolated in the matrix prior to cessation of the orbicular type of crystallization. The presence of nucleus material undoubtedly was a necessary requisite here for the orbicular type of crystallization.

"*Saddle reefs*" *and fracture filling.*—Further evidence of deformation is seen in some parts of the orbicules where differential movement of concentric layers had formed open spaces, especially along the long axis of an ellipsoid. Most of these open spaces were filled by later material, predominantly quartz. A rare open cavity still persists, and in one of these a terminated quartz crystal was observed. The resulting structures are essentially miniature "saddle reefs" (plate 2, fig. 2). In one specimen the "crest" passes into a minute fault which

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Plate 1

Fig. 1. Polished slab of orbicular diorite showing mutual interference of orbicules and ruptured portions of outer orbicule shells isolated in matrix. Note also isolated mafic shell segment around which orbicular type crystallization has continued.

Fig. 2. Polished slab of orbicule showing rupture caused by interference of one structure with a second. Scale in inches.

FIG. 1

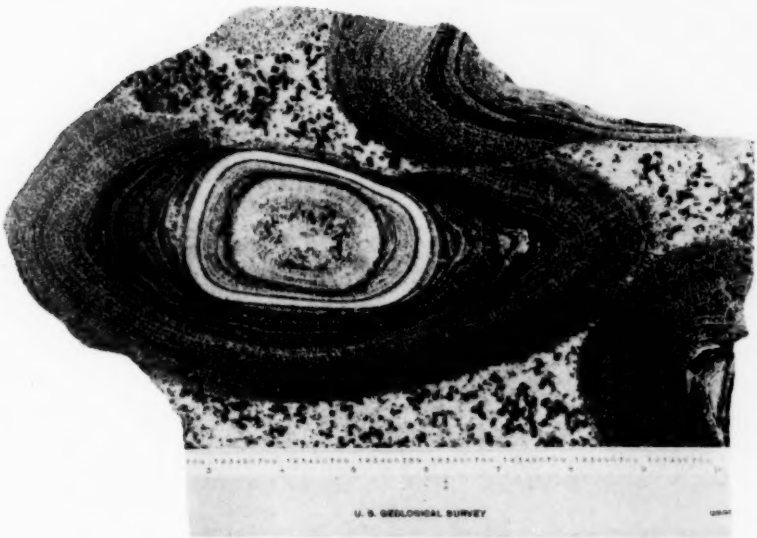

FIG. 2

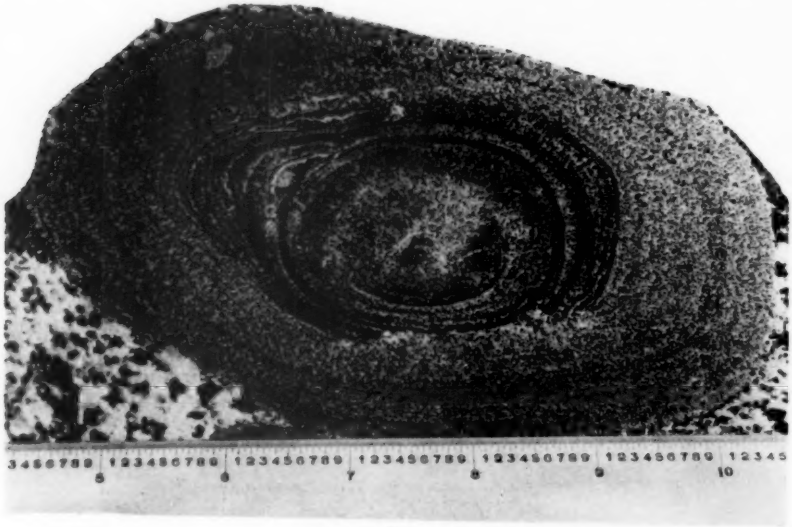

FIG. 1

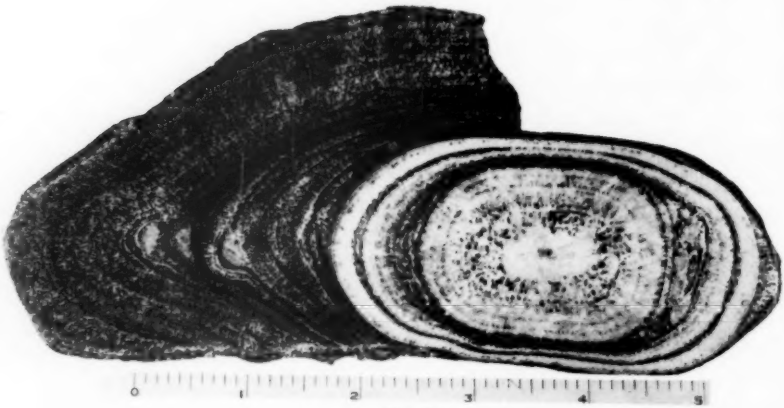

FIG. 2

offsets the outermost concentric bands. Biotite plates which bound the floor and roof of a "saddle-reef" structure may be oriented with cleavage directions nearly at right angles or parallel to the concentric banding. No individual biotite plates were distorted or bent during the period of deformation which formed the reefs, yet concentric bands containing biotite were forced apart mechanically. The evidence suggests that the various concentric mineral bands had some rigidity, but within the shells a degree of plasticity existed which allowed individual minerals to react to the deforming stress without being distorted themselves. The mere thickening and contortion of concentric bands also suggest deformation of a material which was somewhat plastic, whereas the very formation of "saddle reefs" and tearing away of parts of the orbicular structures indicate that it possessed some degree of rigidity.

The quartz filling of "saddle reefs" has been somewhat fractured. In these fractures and along quartz grain boundaries a later mineral of much lower index and relief may be present. It is polysynthetically twinned and probably represents secondary albite. The jagged contacts of the albite(?) skeletal network indicates that part of the action was that of replacement. This albite(?) network generally dies out at the edges of the quartz masses. In only a few places does it clearly penetrate for short distances into minerals bounding the quartz.

Small fractures cutting the orbicules are clearly seen in sawed and polished specimens. These may be continuous from the core to outer rim or may cut only part of the orbicule. They exhibit no pattern. The fractures may be filled with calcite, quartz, epidote, or a combination of calcite-quartz or calcite-epidote. A second generation of minute fractures is generally recognizable only under the microscope. These are usually tight. Offset twin lamellae in plagioclase indicate that there has been a minute amount of movement along some

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Plate 2

Fig. 1. Unpolished orbicule (not cut through nucleus) showing ruptured inner shells around which normal outer shell has been deposited unconformably. Scale in inches.

Fig. 2. Polished section of deformed orbicule showing "saddle-reef" structures filled with later quartz. Scale in inches.

of them. These small second generation fractures in places extend undisturbed into quartz filling of first generation fractures. Rarely veinlets of clear feldspar, probably secondary albite, traverse these fractures through older plagioclase. The younger veinlets may or may not be in optical continuity with the host feldspar.

PEGMATITE AND ITS RELATION TO ORBICULES

Pegmatite associated with the orbicules is composed chiefly of euhedral andesine with lesser quartz, microcline, and orthoclase. Very locally biotite schlieren have been observed. It may be significant that this pegmatite, as well as a second one a mile to the south which contains crude orbicules, are unique in the area. Other pegmatites are common throughout the quartz diorite, but they are granitic and are characterized by a predominance of euhedral microcline crystals admixed with quartz and albite. No orbicules have been found associated with them.

Orbicules are for the most part in a quartz diorite matrix, but a few small orbicules are within somewhat pegmatitic material. It therefore appears that part of the contact of the pegmatite and quartz diorite was the locus of orbicule formation. Although the geologic relationships in the immediate vicinity of the orbicule locality are obscured for the most part by numerous small loose pieces of quartz diorite which are common on most of the slopes of the area, structures in certain loose blocks give further evidence that the orbicules developed along portions of this pegmatite-quartz diorite contact. Quartz diorite blocks containing orbicular structures show a selvedge, in places as much as two inches wide, composed of plagioclase crystals elongated at right angles to the bounding plane. A similar selvedge has also been noted in loose blocks of pegmatitic material.

The pegmatite-quartz diorite contact, aside from being the locus of orbicule distribution, appears to be particularly significant because a comparison of the plagioclase from the pegmatite with that from nuclei of the orbicules shows a striking similarity. Both are about the same size and exhibit multiple twins parallel to the side pinacoid and parallel or nearly so to the basal pinacoid, but more important is the similar anorthite content. Plagioclase from the pegmatite ranges

from about 38 to 42 per cent anorthite; that of the orbicule nuclei ranges from 38 to 41 per cent anorthite. The writer does not consider such a relationship fortuitous. Indeed, weathered surfaces of pegmatite give a strong impression that such plagioclase crystals are incipient orbicules. The occurrence of crude orbicules within a mass of pegmatite at a second locality a mile to the south gives additional support to the idea of a genetic relationship between pegmatite and orbicules. The significance of the pegmatite is further considered below.

CONCLUDING REMARKS

Study of the Alaskan orbicules has revealed several data which are significant in a consideration of their origin. These are:

1. The similarity of mineral composition of the orbicules and surrounding quartz diorite indicates a close genetic relationship, probably crystallization from the same fluid.
2. Orbicules moved around to some degree during and after their formation. They must have been in a fluid or only partly crystalline matrix.
3. Deformation of the orbicules shows that these structures were plastic to a certain extent, inasmuch as relief was largely by flowage, aided possibly by interstitial liquids not yet crystallized.
4. The plastic bodies had some rigidity. This is proved by the "saddle-reef" structures which allowed open spaces to form by differential movement of adjacent shells as a result of deforming stress. It is further corroborated by segments of the outer orbicule shells which have been torn from their original structures and now repose elsewhere in the quartz diorite matrix.
5. The orbicules, which appear to be localized at a contact between andesine pegmatite and quartz diorite, have nuclei composed of large single plagioclase crystals that are remarkably similar in size, anorthite content, and in manner of twinning, to the plagioclase crystals in the pegmatite.

The Alaskan orbicules occur within the quartz diorite batholith and are believed to be primary structures. Ruptured orbicule rims, mutual interference of adjacent orbicules, "saddle-reef" structures, and mineralogic similarity of orbicules

and quartz diorite matrix are logically explained only by a magmatic hypothesis of origin in which the early formed orbicules, somewhat plastic but possessing some rigidity, were free to move in the partly crystallized matrix material. The deformation structures are proof that the quartz diorite matrix could not have been completely solid when the orbicules formed. In this regard it may be noted that quartz, generally considered one of the last constituents to crystallize in an igneous rock, occurs only in the quartz diorite matrix material and as secondary fillings in the orbicule "saddle reefs." The lack of primary quartz in the orbicules thus suggests that they formed prior to the late crystallization of quartz in the matrix material.

The formation of these orbicules is believed to have occupied only part of the cooling period of the quartz diorite host rock (fig. 1). Orbicules began to form sometime after the initial crystallization of the intrusive quartz diorite mass and were completely formed at some time prior to final consolidation of the quartz diorite.

Localization of the Alaskan orbicules near pegmatitic material is thought to be particularly significant, and the following sequence of events is suggested in the evolution of these orbicular structures and the rocks in which they now occur. Closely following the intrusion of the quartz diorite batholith, and while it was yet in a partly unconsolidated but viscous state, a pluglike mass of pegmatite was injected. The intruding pegmatite was also somewhat solidified and contained a number of early formed andesine crystals. That the pegmatitic mass was partly solidified prior to intrusion is supported by the presence of biotite schlieren in places within the pegmatite. Locally along the pegmatite-quartz diorite contact an orbicular type of crystallization began around large andesine crystals. Some of the pegmatite andesine crystals may have become enmeshed in the partly consolidated quartz diorite and become centers of orbicule crystallization. This is suggested by the marked similarity in size, anorthite content, and type of twinning of andesine crystals forming orbicule nuclei to andesine crystals in the pegmatite. A nucleus material of critical size may have been necessary to initiate the orbicular type of crystallization. While the orbicules were forming normal crystallization of material between them proceeded. During

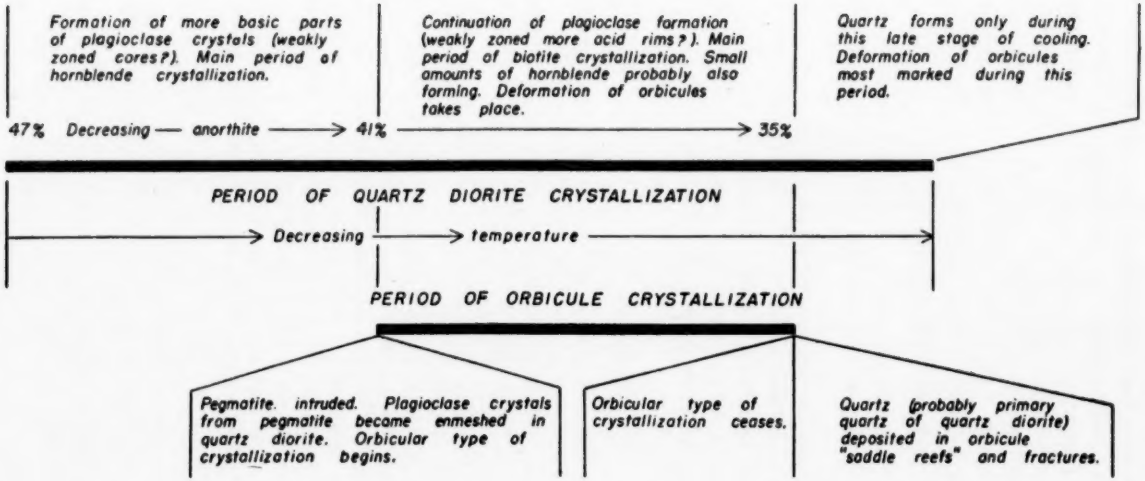
Formation of more basic parts of plagioclase crystals (weakly zoned cores?). Main period of hornblende crystallization.
Continuation of plagioclase formation (weakly zoned more acid rims?). Main period of biotite crystallization. Small amounts of hornblende probably also forming. Deformation of orbicules takes place.
Quartz forms only during this late stage of cooling. Deformation of orbicules most marked during this period.
47% Decreasing — anorthite → 41% → 35%
PERIOD OF QUARTZ DIORITE CRYSTALLIZATION
Decreasing → temperature →
PERIOD OF ORBICULE CRYSTALLIZATION
Pegmatite intruded. Plagioclase crystals from pegmatite became enmeshed in quartz diorite. Orbicular type of crystallization begins.
Orbicular type of crystallization ceases.
Quartz (probably primary quartz of quartz diorite) deposited in orbicule "saddle reefs" and fractures.

DIAGRAM SHOWING SUGGESTED TIME RELATIONSHIP BETWEEN PERIOD OF ORBICULE FORMATION AND PERIOD OF CRYSTALLIZATION OF QUARTZ DIORITE MATRIX

Fig. 1

the crystallization history of the rocks, movement along the pegmatite contact and undoubtedly within the magma chamber continued. As a result the orbicules were deformed at different times during their formation. A minor amount of movement would suffice to cause such deformation. At a late stage in the cooling history of the quartz diorite, the unique conditions resulting in an orbicular type of crystallization, whatever they may have been, ceased to exist, and quartz was precipitated in the normal granular-textured matrix.

ACKNOWLEDGMENTS

The writer is grateful for constructive criticism of the manuscript by various members of the U. S. Geological Survey. Mr. Bernard Wilson checked all index determinations of the plagioclase feldspar and on numerous occasions discussed some of the problems of orbicule genesis with the writer.

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U. S. GEOLOGICAL SURVEY
WASHINGTON, D. C.

ESSAY REVIEW

Structural Geology of North America; by A. J. EARDLEY. Pp. 624, 343 figs., 16 colored plates. Pages 8¼ x 11 inches, elongate left to right, with text in double column. New York, 1951. (Harper and Brothers, Geoscience Series, $12.50).—This unusual volume presents a wealth of information bearing on the structural evolution, since early Paleozoic time, of the North American continent. The author attempts to synthesize much of this information in the form of colored maps, the first of which outlines the major tectonic divisions of the continent. Eleven paleotectonic charts are supplemented by four paleogeologic maps that represent the distribution of rock units in late Devonian, mid-Pennsylvanian, Permian, and late Jurassic times. The tectonic charts, six for the Paleozoic and five for the Mesozoic era, depict for each of the time intervals represented three kinds of major tectonic units: active orogenic belts, areas of epeirogenic uplift, and areas of sedimentary deposition. Sedimentary accumulations no thicker than 1000 feet are represented in white, and those of greater thickness in blue with isopachs at intervals of 5000 feet. This useful device makes the maps tectonically more significant than ordinary paleogeographic maps, which give no indication of variations in thickness. Some details, on both the tectonic and the geologic maps, will be questioned by stratigraphers who are well acquainted with particular areas.

All such maps have their limitations and necessarily involve much speculation. Bold strokes are required in attempting any full picture on a continental scale, and it is easier to criticise or to question one man's offering than to present a more satisfactory substitute. Eardley follows the concept, for which there is a growing body of supporting evidence, that each of the great Paleozoic geosynclines at the borders of the continent was composite, with an outer disturbed belt characterized by volcanism, and an inner belt, generally quieter, in which accumulating sediments were practically free from volcanic products. Each of the outer units was subject to recurrent orogeny, and appears on the maps as an orogenic belt, with progressive changes in outline. Actually the belt along the Pacific coast is shown with little change during the Paleozoic, and in considerable part it is hypothetic, since there is at present no direct evidence for it, during much of the era, from northern California southward. However the characteristic deposits of the inner belt are more continuously displayed and by analogy furnish an argument for continuity of the outer belt as well. Generous extrapolation and assumption are required for full representation of the belt along the Atlantic seaboard also, especially for certain periods.

The following random comments on the maps are samples of many evoked by close study. The "Taconic" belt shown on plate 2

is so nearly identical with the "Taconic and Acadian" belt of plate 3 that students may well wonder whether orogeny was continuous, within nearly fixed limits, from Ordovician through Devonian time. In New England and the Maritime provinces, however, where effects of the Taconian and Acadian orogenies are best displayed, there seems to be some geographic distinction. On plate 5, representation of Mississippian deposits in the Cordilleran belt will mystify readers until an explanation of a pattern is supplied. The distribution of orogenic activity suggested for the Appalachian region in Permian time is unlikely to meet wide concurrence. Probably every student of Cordilleran geology will wonder at the omission of a colored map to represent the climactic disturbance of the Rocky Mountain belt—Eardley's "Laramide orogeny"—which however is outlined in figures 162 and 176. These and many other questions are of course no condemnation of the maps, which doubtless will serve as a basis for many fruitful discussions.

The text is far too extensive and detailed for any adequate appraisal in a short review. After a brief resumé based on the set of colored maps, the regional treatment starts logically with the great "stable interior" of the continent and proceeds to analysis of the more eventful histories of the marginal belts. A chapter entitled "Paleozoic Cordilleran Geosyncline" was published separately in 1947 and was greeted with favorable comment as an excellent digest of a voluminous literature, and a good presentation of stimulating concepts. Instead of continuing directly with later development of the Cordilleran region, the author shifts the scene to "Appalachian Mountain Systems," probably with the thought of clearing up outstanding Paleozoic events before introducing the major activities of later periods. There is no discussion of the Appalachian geosyncline, analogous to that for the western belt; description and analysis of existing structural features are given for successive geographic units, from southeastern United States to Newfoundland, with emphasis on "phases" of deformation. The general plan of historical sequence is compromised, for convenience of treatment, and chapters on the Triassic basins and the Atlantic Coastal Plain are included with those devoted to Paleozoic structure. From the Appalachians the reader is led to the Ouachita and Marathon deformed belts, and to the related Coahuila structures of Mexico. Consideration of the Wichita unit and the "Ancestral Rockies" leads logically to the Mesozoic and Cenozoic deformations in the vast Cordilleran region, to which are devoted about half of the chapters and fully half of the total text. Alaska is treated separately, though with recognition of its place in the major Cordilleran unit. There follows a chapter on Arctica, and the final sections of the book deal with the Gulf Coastal Plain and the Antillean-

Caribbean region. Thus all major parts of the continent receive attention.

Doubtless every student of tectonics will find in the book much of constructive value, and also, in sections that treat districts or structural units with which the student has close acquaintance, some points open to question or to direct criticism. In addition to assembling an unusual array of descriptive and factual matter, the author has gone into outstanding problems and controversies, such as the extent and relative importance of successive orogenies in the Appalachian region, the age of the Glenarm Series, the date and the structural setting of the Idaho batholith. Study of these extensive and varied materials is rewarding, perhaps all the more because of disagreements with certain of the author's methods and conclusions. The use of some important terms, defined on page 11 as distinctive and not interchangeable, seems confused in some passages. Taconic and Acadian *systems* are described for New England (Chapter 10), but elsewhere (e.g., pp. 183, 200-201) we read of Taconic and Acadian *orogenies*. The three provinces of the southern Appalachians—Valley and Ridge, Blue Ridge, and Piedmont—are referred to (p. 69) as "structural systems," although in the book generally the term *structural system* is applied to a more or less self-contained geographic and structural unit (e.g., Ouachita, Marathon, Wichita systems). Using one and the same term for different features or concepts causes ambiguity, as on page 152 where "Logan's line," commonly understood as marking a thrust front in the Champlain Valley, is identified as "the trace of the Taconic thrust . . . ." Since the term *Taconic* is firmly attached to a specific thrust in central Vermont, there is much to be said for adoption of *Taconian* for the Ordovician orogeny. Similarly, attaching the term "Rockies" to every unit of the "Laramide" belt, including those in southern Arizona and in Mexico (fig. 162) seems unfortunate, since it violates well established usage for the Rocky Mountains and their subdivisions.

Names and definitions for orogenic events in the Cordilleran region present increasing difficulty with growth of knowledge. The Nevadan orogeny, in Eardley's usage, is confined to late Jurassic time (p. 246), and the Laramide orogeny occurred "in very late Cretaceous, Paleocene, and Eocene time" (p. 284). Other pulses of deformation that intervened, in time and space, are recognized, and two of these are represented in plates 15 and 16. However, the map (fig. 176) appears rigid in its division of the wide Cordilleran area into Nevadan and Laramide belts, and many students will take this at face value regardless of explanations in the legend and the text. As W. M. Davis once commented, "We suffer more from our diagrams than from our ideas."

Proofreading for the first edition of such a volume is an onerous task, and it is not surprising to find numerous errors. Insofar as these are merely typographic, or involve single words, they can be rectified without difficulty in a future printing. Wrong sources are cited for some diagrams, and a number of references given in the text do not appear in the bibliography. More serious is the evident confusion of the author regarding passages from literature on the Hope fault in Idaho (pp. 312-313); two quotations supposed to demonstrate a contradiction actually are entirely consistent.

The line drawings are an outstanding feature of the book. Many are the author's own work; others are reproduced from other sources, either directly or with modification. Almost without exception they are well executed, with lines and letters clear to the unaided eye. The text is in pleasing and legible type. Doubtless there will be some dissatisfaction with the format of the book—it is too long to fit into an ordinary book shelf. Binding in quarto form, similar to the volumes of van Bemmelen's "The Geology of Indonesia" (The Hague, 1949), would have avoided this difficulty, though admittedly the long horizontal page makes for convenience in consulting most of the maps and diagrams while reading the text. Registration of colors on maps is not uniformly good.

The editor and publishers announce the book as "the first in any language which describes in some detail the structural evolution of an area as large as a continent." This statement is hardly fair to von Bubnoff's *Geologie von Europa* (Gebrüder Bornträger, Berlin, 1926-36, 1603 pp., 245 figs., several folded plates), and to Krenkel's *Geologie Afrikas* (Gebrüder Bornträger, 1925-38, 1918 pp., 412 figs., several folded plates). Although the titles of these works do not so announce, the tectonic evolution of each continent concerned is emphasized, and much detail is presented in both structure and stratigraphy. Eardley's volume is a pioneer for North America.

The book is intended primarily for use by advanced undergraduates. Surely it contains much that serious college students will find valuable and stimulating; but probably a considerable reduction of detail and a greater degree of synthesis would make a more practical volume for use in the average undergraduate course. The book as it stands has abundant materials for seminar discussion, and will be a valuable source of information for all professional geologists. The bibliography, containing about 700 titles, is a large asset for all users.

Authorship of *Structural Geology of North America* required much courage and enormous labor. Any receptive user of the book, whether student or mature geologist, will gain from it useful information and valuable concepts. Professor Eardley has earned our gratitude.

CHESTER R. LONGWELL

REVIEWS

Métamorphisme, Silicifications et Pédogénèse en Bohême Méridionale; by P. ELLENBERGER *et al.* Pp. 189, illustrated. Annales scientifiques de Franché-Comte, troisième Année. Université de Besançon, 1948.—This is not a review, for it would be presumptuous for a paleontologist to review a book dealing mainly with petrology. Let us rather call it a notice, which is being written because of the subtitle of this volume: *Travaux pétrographiques et paléobotaniques de l'Université de captivité d'Edelbach* (*Oflag XVII A* — 1940-1945).

Part one deals with the sub-soil and its history, and consists of two chapters, on the consequences of a syntectonic basic intrusion, and on the process of metasomatic silicification. From their studies the authors develop theories which they express in a table of mineral evolution. Part two is less homogeneous in material. There are notes on a fragmentary belemnite, on the composition of local gravels, on fragments of silicified woods, and theoretical considerations based on them. Briefly the authors record the discovery of archeological remains. They note the weathering and erosive processes seen at work during their study. And finally, a section is devoted to the plants of the region, with careful records of those which increased in abundance during the four years, those which became scarcer, those which were unchanged in numbers, with suggestions on the causes of these differences in behaviour.

This, then, is an account of the geology of a plot 400 meters square, as it was studied by a group of men who lived there for four years, gave each other courses in geology, and organized a "Cercle de Géologie." It is an acconut of how they studied the available rocks with a microscope in which the polariser was a pile of cover slips; of how they cemented their slides to pieces of glass with a mixture of violin resin and cooking oil. How, as each tedious day passed, they carefully recorded observations, how they organized seminars, and how new theories developed from their deliberations. All this while they were far from libraries or laboratories, with only the recollections of the group to serve as background for their study. It is an account of how the weary burdened men, free at last, gladly added to their loads their precious specimens, carrying them in their packs on the long march homeward.

There is no plea for pity in this book. The authors point out that Edelbach was not a death camp, that the prisoners were allowed to receive some packages from France, and thus obtain some supplies and a very few textbooks, which were learned by heart. But 400 metres is about one quarter of a mile, and hundreds of men lived there for four years.

I think this book will appear regularly on my reading lists (silicification is a legitimate interest of even a paleontologist) so that my students may meet it, and meeting it be proud of the Science they are entering. And that they may see, and think long about, such a picture as plate 2, figure 1: a simple ink sketch of a peneplane running across the country with the hills beyond and beyond, and, in the lower left hand corner of the sketch, unobtrusively, the barbed wire. G. WINSTON SINCLAIR

A Palaeogeographical Atlas of the British Isles and Adjacent Parts of Europe; by LEONARD J. WILLS. Pp. 64; 22 plates, 4 figs., 11 x 8½″. London and Glasgow, 1951 (Blackie and Son, Ltd., 66 Chandos Place, London, and 17 Stanhope Street, Glasgow, 21 shillings).—The title-page of this atlas bears the phrase "Geology the Geography of the Past." The attitude to geological data implied in this definition is one that has characterized Professor Wills' work. His distinguished contributions to British Palaeozoic and early Mesozoic stratigraphy and palaeontology, and to Pleistocene geology, give him a rare understanding with which to embark on the preparation of this atlas. It is no mere compilation and presentation of known facts, but an interpretative synthesis of data. No atlas of this type for Great Britain has previously been attempted, though many palaeogeographical maps have appeared through the years. A comparison with earlier maps reveals not only the immense increase in amount of data available in the past fifteen years, but also the originality of many proposed solutions for certain problems, and the value of keeping ever in mind the close relationship with the European continent. For example, in departing from the traditional classification of the largely unfossiliferous Permian and Triassic strata, Professor Wills has been enabled to give a new conception of the changing geographies of these periods. In this and in many other ways, the author has re-interpreted the data, and his views will suggest fresh approaches to many problems. Thirty-two palaeogeographical maps are drawn upon the present-day outline of the British Isles, at least two for each system, with a maximum of seven devoted to the Carboniferous. Only the Cambrian is omitted, and this because of the limited evidence available. The use of two tones of overprint, and a wide variety of shading and symbols, enables the portrayal not only of the distribution of land and marine, brackish and fresh waters, but also the known extent of different facies, the occurrences of volcanic rocks, areas where the base of the series or stage portrayed is transgressive, etc. Present outcrop belts are rarely shown. Great attention has been paid to the presentation of structural features,

and their relation to areas of sedimentation and past geography. Beginning with main structural units of the present, Caledonian tectonics and pre-Mesozoic structures are dealt with in their appropriate place. Together with the maps of the British Isles, 24 palaeogeographical and tectonic maps of Europe are given, so that at each stage the palaeogeography and structure of Britain may be related to that of the adjacent continent. A brief description accompanies each map, and a short bibliography is appended. It was the aim of the author that the publication of this atlas should be at a price within the reach of the British student, and this objective has been attained, and a high standard maintained in the production. Though some of the maps of Europe are small, they are adequate to fulfill their intention of portraying broad features.

Professor Wills warns in his preface of the "great risk, as I see it, . . . that the reader may be misled into thinking that each or any of these maps . . . gives an accurate picture of the geography of the period which it is designed to illustrate." Nevertheless, this atlas will be an indispensable aid in gaining an understanding of British stratigraphy, and especially for students of other countries. It may be regarded as the result of prolonged reflection on the subject-matter, and as having taken full account of the views of others. Many stratigraphical terms, names of faults, ancient land masses, orogenic phases, etc., the meaning of which is frequently unknown to foreign readers of British writings, are here explained and located. Since present outcrops of strata are rarely indicated, a geological map will also be essential for the reader, and a reference might have been included to the twenty-five miles to one inch map of Great Britain, or to the ten mile map, which does not include Ireland (both are published by the Geological Survey of Great Britain). Apart from helping to comprehend the details of British stratigraphy, many lessons of more general application may be drawn from these leaves. For example, the portrayal of the lower Palaeozoic geosyncline, the succeeding orogen, and the geography of this area in later periods, offers a well-documented story of the distribution of sedimentary facies, the inter-relation between these facies and tectonics, and the subsequent behavior of an intensely folded region.

H. B. WHITTINGTON

Josiah Willard Gibbs; by LYNDE PHELPS WHEELER. Pp. viii, 264, illustrated. New Haven, 1951 (Yale University Press, $4.00). —Because of the great originality and generality of his contributions, it is universally agreed that Josiah Willard Gibbs is to

be classed as one of the greatest scientists of modern times and possibly the greatest scientist the United States has yet produced. Up to the present, scientists who knew Gibbs and who from their own training and knowledge could properly evaluate his work have written only biographical sketches of this interesting figure. Some of these sketches and reviews, for example those by H. A. Bumstead, E. B. Wilson and C. S. Hastings, are excellent. And of course there is the comprehensive, two-volume *Commentary* on Gibbs' scientific writings, edited by F. G. Donnan and A. Haas. There has been a need, however, of a full-length biography such as Lynde Wheeler has now written, appraising Gibbs' life and work in some detail.

Only a scientist with a facile pen and with full knowledge of the complexities of Gibbs' contributions could write this biography. Lynde Wheeler, who was a student of Gibbs and later taught at Yale the Gibbsian thermodynamics and statistical mechanics, has produced a fascinating account of Gibbs' life and a non-mathematical description of his accomplishments which is technically correct and yet not beyond the comprehesion of the general reader. It is not easy to explain Gibbs' results to the non-scientist! The author has had access to family documents and much unpublished material which enables him to give convincing explanations of the main decisions made by Gibbs during his career. Wherever possible Wheeler documents his statements, while on points for which written evidence is missing his speculations are always plausible because they are based on full knowledge of the developments in physical science and mathematics in the last three decades of the nineteenth century.

We learn that Gibbs was by no means a recluse. From the beginning he had a strong interest in applied mechanics. Early in his career he invented a "center vent" hydraulic turbine, a railway car brake, and an improved steam-engine governor. He corresponded with numerous scientists in this country and Europe, and he sent reprints of his research papers to a long list of names. The importance of his work was fully recognized in his lifetime by men like Lord Kelvin, Ostwald, Maxwell and J. J. Thomson. Just one year after publication of the famous monograph "On the Equilibrium of Heterogeneous Substances," when he was barely forty years old, he was elected to the National Academy of Sciences. From then to the time of his death in 1903 many honors came to him.

It is true, however, that Gibbs was no "joiner." Nor did he ever attend many meetings of societies outside New Haven. He simply did not want to be distracted from his work by such contacts.

During all his years on the Yale faculty Gibbs taught probably fewer than one hundred students. And those few who did their doctoral dissertation under his direction were for the most part "on their own."

Gibbs led at Yale the quiet, scholarly life among pleasant, congenial surroundings which brought him complete happiness. This explains why he declined an offer from Johns Hopkins at a salary of $3,000 per year to remain at Yale on a $2,000 salary. Money meant little to him in comparison with the deep satisfactions to be derived from the production of important scholarly research in a family group and among friends he loved. All remember Gibbs as a cheerful, sociable, serene man, sure of himself and of the correctness and importance of his scholarship.

There are interesting items in the appendices: a genealogical chart, Gibbs' first paper read before the Connecticut Academy in 1866, a catalogue of Gibbs' "Scientific Correspondence," Gibbs' mailing lists for reprints, and a complete bibliography of articles and books about Gibbs. The index is excellent.

Dr. Wheeler has given us an extremely interesting and readable biography of one of the most original scholars of modern times. As Yale University now prepares to celebrate the two hundred and fifty years of its service to mankind, it is fitting that there appears this tribute to its greatest professor. W. W. WATSON

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